Culprit of the Eastern Pacific Double-ITCZ Bias in the NCAR CESM1.2

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ABSTRACT

The eastern Pacific double-ITCZ bias has long been attributed to the warm bias of SST in the southeastern Pacific and associated local air–sea interaction. In this study, we conducted two simulations using the NCAR CESM1.2.1 to demonstrate that significant double-ITCZ bias can still form in the eastern Pacific through air–sea coupled feedback even when there is cold SST bias in the southeastern Pacific, indicating that other nonlocal culprits and mechanisms should be responsible for the double-ITCZ bias in the eastern Pacific. Further analyses show that the oversimulated convection in the northern ITCZ region and Central America in boreal winter may result in biases in the surface wind fields in the tropical northeastern Pacific in the atmospheric model, which favor the cooling of the ocean mixed layer through enhancement of latent heat flux and Ekman upwelling. These biases are passed into the ocean model in coupled simulations and result in a severe cold bias of SST in the northern ITCZ region. The overly cold SST bias persists in the subsequent spring, leading to the suppression of convection in the northern ITCZ region. The enhanced low-level cross-equatorial northerly wind strengthens the wind convergence south of the equator and transports abundant water vapor to the convergence zone, strengthening the southern ITCZ convection. All these processes lead to the disappearance of the northern ITCZ and the enhancement of the southern ITCZ in boreal spring, forming a seasonally alternating double-ITCZ bias. This study suggests that convection biases in the northern ITCZ region and Central America in boreal winter may be a culprit for the double-ITCZ bias in the eastern Pacific.

1. Introduction

The double–intertropical convergence zone (ITCZ) bias is a long-standing tropical bias that has plagued generations of coupled general circulation models (GCMs) for more than two decades (Mechose et al. 1995; Lin 2007; Zhang et al. 2015). The annual mean precipitation simulated by most coupled GCMs tends to exhibit two parallel zonal belts of maximum precipitation straddling the equator over the central and eastern Pacific Ocean. However, the observed annual mean precipitation shows only one zonal ITCZ rainband north of the equator and a southeastward South Pacific convergence zone (SPCZ) extending from eastern Papua New Guinea toward French Polynesia, although two weak zonal ITCZ precipitation bands, referred to respectively as the northern and southern ITCZs, occur during boreal spring in observations (Zhang 2001). The annual mean precipitation bias in the southern ITCZ region is commonly accompanied by a warm bias of sea surface temperature (SST) and anomalous convergence of wind

stress (Zhang et al. 2007; Song and Zhang 2009; Oueslati and Bellon 2015).

The double-ITCZ bias has distinct regional and seasonal characteristics (e.g., de Szoeke and Xie 2008; Oueslati and Bellon 2015; Adam et al. 2018). In the central Pacific, coupled GCMs tend to overestimate the precipitation in the southern ITCZ region and hence produce double-ITCZ rainbands year-round. However, in the eastern Pacific, coupled GCMs tend to simulate a seasonally alternating ITCZ with an overestimated precipitation band south of the equator during the boreal winter and spring and north of equator during summer and autumn, still resulting in a double ITCZ in annual climatology.

GCMs forced with observed SST commonly exhibit a very weak double-ITCZ bias, which is severely exacerbated in corresponding coupled GCMs, indicating that air–sea interaction and simulated SST bias play important roles in the formation of double-ITCZ bias (Lin 2007; Oueslati and Bellon 2015; Zhang and Song 2010). In the eastern Pacific, the double-ITCZ bias has long been believed to be associated with local air–sea interaction. The overestimated and long-lasting

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precipitation band in the southern ITCZ region is historically attributed to the warm bias of SST in the southeastern Pacific (SEP). Several studies (Ma et al. 1996; Yu and Mechoso 1999; Dai et al. 2003) found that the underestimated stratus cloud cover off the western coast of South America allows more downward short-wave radiative flux to warm the local SST. The westward extension of this warm bias may result in anomalous warm SST in southern ITCZ region. Artificially increasing the stratus cloud fraction off the western coast of South America (Ma et al. 1996; Yu and Mechoso 1999) or improving the stratus cloud parameterization (Dai et al. 2003) actually significantly alleviated the warm SST bias and hence the double-ITCZ bias in the SEP. However, the extensive bias of the equatorial cold tongue was exacerbated in those studies. Zheng et al. (2011) found pronounced warm SST biases in the SEP stratus region in all coupled GCMs participating in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4). Their detailed upper-ocean heat budget analyses showed that the surface heat fluxes bias cannot explain the warm biases of SST, while the biases in ocean heat transport by Ekman currents largely contribute to the warm biases of SST. Wang et al. (2015) investigated the double-ITCZ bias in the eastern Pacific in the NCAR Community Earth System Model, version 1 (CESM1), and attributed the bias to a positive local air–sea feedback. They argued that the overly weak surface easterly wind in the SEP results in eastward ocean current around 5°S and less evaporation. The eastward warm ocean advection and reduced latent heat flux lead to warm SST bias that favors convection development, which further reduces the wind speed at the new convergence zone, forming a positive feedback. However, the processes through which the ocean current is changed from westward to eastward around 5°S by overly weak surface easterly wind were not well explained in their study. The quantitative contributions of ocean zonal temperature advection, surface latent heat flux, and other ocean processes were not evaluated either. To date, there is still no satisfactory theory that can explain the formation mechanism of double-ITCZ bias in coupled GCMs in the eastern Pacific.

Instead of identifying the causes for warm SST bias in the SEP, Song and Zhang (2016) recently directly investigated the impacts of warm SST biases in the SEP on the double-ITCZ bias in the NCAR CESM1 by prescribing the SST in the SEP (covering 5°–25°S and east of 120°W) with observations. The results show that even though the SST in the SEP is constrained by observations, a significant precipitation bias in the southern ITCZ region still persists during February–

June although the bias is reduced somewhat comparing to fully coupled simulation. This indicates that the spring southern ITCZ bias over the SEP is not entirely controlled by local SST bias. Other unknown culprits still exist and need to be identified.

In this study, we conducted a set of experiments using the NCAR CESM1.2.1 and demonstrated that the significant double-ITCZ bias can still form in the eastern Pacific through air–sea coupled feedback even when there is cold SST bias in the SEP. The results indicate that warm SST bias in the SEP may not be the primary culprit for the eastern Pacific double-ITCZ bias, although it may amplify the bias. An unknown nonlocal feedback mechanism should be responsible for the double-ITCZ bias in the eastern Pacific. The further analyses suggest that excessive convection in the northern ITCZ region and Central America in boreal winter may result in biases in surface wind fields in tropical northeastern Pacific in the atmospheric model, which favor the cooling of ocean mixed layer through the enhancement of latent heat flux and Ekman upwelling. These biases are passed into the ocean model in coupled simulation and result in severe cold bias of SST in the northern ITCZ region. The overly cold SST bias persists in the subsequent spring, leading to the suppression of convection in the northern ITCZ region. The enhanced low-level cross-equatorial northerly wind strengthens the wind convergence south of the equator and transports abundant water vapor to the convergence zone, strengthening the southern ITCZ convection. All these processes lead to the disappearance of the northern ITCZ and the enhancement of the southern ITCZ in boreal spring, forming a seasonally alternating double-ITCZ bias.

The rest of this paper is organized as follows. Section 2 briefly describes the model, experiments, and observational datasets. Section 3 identifies the double-ITCZ bias and a possible culprit, and discusses the formation mechanism of double-ITCZ bias in the eastern Pacific. Section 4 presents a summary of results and conclusions.

2. Model, experiments, and data

a. Model

The model used in this study is the NCAR Community Earth System Model, version 1.2.1 (CESM1.2.1), which consists of seven component models (atmosphere, sea ice, land, river runoff, ocean, land ice, and ocean wave) linked by a coupler. The atmosphere model component, the Community Atmosphere Model, version 5.3 (CAM5.3), is a global atmospheric general circulation model with 30 levels in the vertical. The

default CAM5 physics package and finite-volume (FV) dynamical core at $1.9^\circ \times 2.5^\circ$ resolution (lat $\times$ lon) are employed for the atmosphere model in this study. The ocean model is an extension of the Parallel Ocean Program (POP), version 2, from Los Alamos National Laboratory (LANL). Version 6 of the 1° Greenland Pole ocean–ice grid is used for the active ocean and sea ice models. The Community Land Model, version 4.0 (CLM4.0), with carbon and nitrogen (CN) at $1.9^\circ \times 2.5^\circ$ resolution (lat $\times$ lon) is used for the land model. (For more details of the model, refer to <http://www.cesm.ucar.edu/models/cesm1.2/>.)

b. Experiments

Both the fully coupled simulation and the Atmospheric Model Intercomparison Project (AMIP) simulation are conducted using standard CESM1.2.1 in this study. The AMIP simulation that replaces the active ocean and sea ice models in the CESM1.2.1 with data from the Hadley Centre Global Sea Ice and Sea Surface Temperature dataset (HadISST) is conducted from January 1979 to December 1987. The 5-yr averages for years 1983–87 are used in the analysis, which is referred to as AMIP. The 11-yr fully coupled simulation employs the B1850C5CN component setting configuration and f19_g16 grid configuration. That means the simulation is performed with preindustrial forcing at the 1850 level and the atmosphere model component uses the same grid resolution as in the AMIP simulation. The coupled simulation is a hybrid run and is initialized with the datasets at year 70 of the NCAR preindustrial control run (reference case: b40_1850_2d_r07c5cn_160jp). Note that the formation of a double-ITCZ bias is a very quick process in coupled GCMs. Liu et al. (2012) found that the double-ITCZ bias develops within the first 2 years of simulation, and after 5 years the bias is almost the same as that in the long-term climatological simulations. Therefore, the 5-yr averages for years 7–11 of coupled simulation are used in the following analysis, which is referred to as CPL.

Note that very similar double-ITCZ bias exists in both coupled preindustrial and twentieth-century simulations of CESM1.2.1 (Song and Zhang 2016), indicating that the greenhouse gases level has very little impact on the double-ITCZ biases in coupled GCMs. Thus, the major difference in the double-ITCZ bias between the AMIP and coupled simulation comes from the air–sea interaction and associated SST changes. The coupled simulation is evaluated with observations to identify the double-ITCZ bias and its formation mechanism. By comparing errors in the coupled simulation and the corresponding AMIP simulation, we can examine to what extent the errors can be traced

back to the atmospheric model and to what extent the errors can be attributed to the atmosphere–ocean feedback. Note that the coupled simulation with preindustrial forcing produces colder SSTs relative to the AMIP simulation in the SEP, yet it still produces an enhanced double-ITCZ bias, which inspires us to identify the new culprit and mechanism for the eastern Pacific double-ITCZ bias.

c. Data

The observational data used in this study include the monthly mean surface latent heat flux, sensible heat flux, shortwave radiative flux, longwave radiative flux, SST, specific humidity at 2-m height, and daily zonal and meridional wind at 10-m height from the National Centers for Environmental Prediction (NCEP)–Department of Energy (DOE) AMIP-II Reanalysis (NCEP atmospheric reanalysis 2; Kanamitsu et al. 2002) for years 1983–87, the monthly mean ocean temperature, mixed layer depth, total downward heat flux at the surface, wind stress, and current velocity data of the NCEP Global Ocean Data Assimilation System (GODAS) (Ji et al. 1995; Behringer et al. 1998) from 1983 to 1987, and SST from HadISST (Rayner et al. 2003) for years 1983–87. The monthly mean meridional wind speed and pressure vertical velocity from the ECMWF interim reanalysis (ERA-Interim; Simmons et al. 2007) between 1989 and 2005, and the Global Precipitation Climatology Project (GPCP) monthly precipitation analysis (Adler et al. 2003) for years 1979–2009 are also used in this study. Note that the NCEP GODAS is forced by surface heat flux, momentum flux, and freshwater flux from the NCEP–DOE AMIP-II Reanalysis (<https://www.esrl.noaa.gov/psd/data/gridded/data.godas.html>). However, it provides only the total instead of each component (latent heat, sensible heat, shortwave radiative, and longwave radiative) of surface heat flux. Therefore, the surface heat flux of NCEP–DOE AMIP-II Reanalysis is used to identify the impact of each component of surface heat flux on ocean mixed layer temperature, and its shortwave radiative flux component is used to determine shortwave radiation penetration in mixed layer for the budget analysis of GODAS.

3. Results

Figure 1 presents the seasonal cycle of monthly SST averaged between 90° and 120° W from the HadISST observation, CPL simulation, and their difference. Since the AMIP simulation is forced by the SST data from the HadISST, the difference between HadISST and CPL also represents the SST changes due to the air–sea interaction in the CESM1.2.1. The seasonal cycle of

FIG. 1. Seasonal cycle of SST ($^{\circ}\text{C}$) averaged over 90° – 120°W for (a) HadISST and (b) CPL and (c) the difference between CPL and HadISST.

observed SST (Fig. 1a) shows pronounced asymmetry about the equator. The observed SST in the northern ITCZ region is warmer than 27°C throughout the year and increases to higher than 28°C in March, and reaches its maximum (about 29°C) in June. Since then the SST gradually decreases to its minimum (about 27°C) in January and February. In contrast, the SST in the southern ITCZ region is warmer than 27°C only from March to May with maximum up to 28°C , which is slightly cooler than its northern counterpart. The CPL simulation generally reproduces the observed SST seasonal cycle with comparable maximum (about 29°C) occurring in September in the northern ITCZ region. A striking difference between the observed and simulated SST is that the simulated SST decreases sharply from October (about 29°C) to January ($<26^{\circ}\text{C}$) and then is maintained at its minimum level ($<26^{\circ}\text{C}$) till March. This produces a significant cold SST bias up to 2.5°C in the northern ITCZ region from November to June.

FIG. 2. Seasonal cycle of precipitation (mm day^{-1}) averaged over 90° – 120°W for (a) GPCP, (b) AMIP, and (c) CPL.

Another notable change is that the coupled simulation does not produce a warm SST bias relative to the AMIP simulation in the southern ITCZ region; actually it is slightly colder relative to the AMIP simulation. Since the double-ITCZ bias in the eastern Pacific is historically attributed to warm SST biases in the SEP, it is interesting to see if the double-ITCZ bias is still exacerbated in the coupled simulation with cold SST bias in the SEP.

Figure 2 presents the seasonal cycle of monthly precipitation averaged between 90° and 120°W from the GPCP observation and AMIP and CPL simulations. The seasonal cycle of GPCP precipitation shows similar characteristics to that of observed SST, indicating the modulation of precipitation by SST. The precipitation greater than 3 mm day^{-1} persists year-round in the northern ITCZ region. Following the increase of SST, precipitation in the northern ITCZ region gradually increases and extends northward from April to its maximum in July. After September, precipitation

gradually decreases and retreats back to 5°N in January. In contrast, precipitation greater than 3 mm day^{-1} in the southern ITCZ region only occurs in March and April with maximum less than 5 mm day^{-1} , when the underlying SST is warmer than 27.5°C (Fig. 1a). The southern ITCZ precipitation is comparable to its northern counterpart, resulting in a weak double ITCZ in boreal spring. In the AMIP simulation, precipitation is systematically oversimulated. The northern ITCZ precipitation is greater than 5 mm day^{-1} throughout the year. The precipitation greater than 3 mm day^{-1} in the southern ITCZ region persists from January to June. The maximum precipitation up to 8 mm day^{-1} in the southern ITCZ region occurs in March and April, which is still comparable to its northern counterpart, resulting in a stronger double ITCZ in boreal spring. One notable difference between the AMIP and GPCP precipitation is that the simulated precipitation in the northern ITCZ region is not seasonally decreased from October to December, leading to the largest positive precipitation bias in boreal winter. The air–sea coupling dramatically influences the precipitation distribution in the CPL simulation. The northern ITCZ precipitation almost disappears from January to May and the southern ITCZ precipitation is significantly increased with a maximum up to 12 mm day^{-1} from January to June, resulting in pronounced seasonally alternating type (Mechose et al. 1995) of double-ITCZ bias. This result shows that significant double-ITCZ bias can still form in the eastern Pacific through air–sea coupled feedback even when there is cold SST bias in the SEP, indicating that this double-ITCZ bias in the eastern Pacific should not be caused by changes in local SST forcing in the SEP; other nonlocal culprits and mechanisms should be responsible for the double-ITCZ bias in the eastern Pacific.

To further look into the double-ITCZ bias, Fig. 3 shows the meridional distribution of surface precipitation and SST averaged between 90° and 120°W for March, when the errors are most significant. The precipitation distribution of GPCP (Fig. 3a) shows two weak peaks at 4.5 mm day^{-1} located around 5°S and 5°N , representing the southern and northern ITCZ respectively. The AMIP simulation also produces two precipitation peaks around 5°S and 5°N but with much higher intensity (about 7.5 mm day^{-1}). This positive precipitation bias comes from the atmospheric model response to observed SST. When the SST is simulated by the ocean model, the precipitation distribution is changed dramatically. The peak of the northern ITCZ disappears and the peak of southern ITCZ is increased to 11.5 mm day^{-1} around 8°S . The SST distribution (Fig. 3b) shows that the simulated SST by the coupled model is colder than that of AMIP simulation by more

FIG. 3. Meridional distribution of (a) surface precipitation and (b) SST averaged over 90° – 120°W for March.

than 2°C in the northern ITCZ region, by more than 3°C in the cold tongue near the equator, and by about 0.5°C around 5°S . The SST in the coupled simulation is even colder than 26°C between 5° and 12°N . Note that studies of the relationship between SST and convection (e.g., Zhang 1993) showed that deep convection is weak and rarely observed where SST is colder than 26°C in the tropics. The deep convection tends to be more intense as SST increases from 26° to 29°C (Waliser et al. 1993). This indicates that the cold SST bias produced by the coupled model in the northern ITCZ region may suppress deep convection north of the equator. Figure 4 presents the meridional–vertical distribution of vertical velocity and wind vectors in the atmosphere averaged between 90° and 120°W for March from ERA-Interim and the AMIP and CPL simulations. Corresponding to two observed weak precipitation peaks in Fig. 3, ERA-Interim shows two shallow centers of upward motion around 5°S and 5°N (Fig. 4a), respectively. Coinciding with the stronger precipitation peaks in the AMIP simulation, the two regions of ascending motion are much deeper in the AMIP simulation (Fig. 4b). When SST is simulated by coupled model, the overly cold SST in the northern ITCZ region suppresses the convection, resulting in the disappearance of upward motion in the northern ITCZ region in the CPL simulation (Fig. 4c). The enhanced

FIG. 4. Meridional-vertical distribution of vertical velocity (shading; hPa day^{-1}) and wind vectors (m s^{-1}) averaged over 90° – 120°W for March from (a) ERA-Interim, (b) AMIP, and (c) CPL.

low-level cross-equatorial northerly wind (Fig. 4c) strengthens the convergence south of the equator and transports abundant water vapor to the convergence zone, which fuels the southern ITCZ convection. In addition, feedback processes may further strengthen the southern ITCZ. Figure 4c shows that southern ITCZ convection produces descending motion over the northern ITCZ region, which can be linked to low-level equatorward flow through Sverdrup balance (Takahashi and Battisti 2007b). This subsidence may further suppress convection in the northern ITCZ region and the associated low-level equatorward flow may transport more water vapor to the southern ITCZ region, which reinforces the southern ITCZ. To

demonstrate the role of moisture transport, we calculated the water vapor budget of CPL and AMIP simulations in the southern ITCZ region for March. The average precipitation from CPL over 5° – 10°S , 90° – 120°W is $11.36 \text{ mm day}^{-1}$, and the contribution from ocean surface evaporation is 2.69 mm day^{-1} . The contribution from moisture flux convergence below 850 hPa is $10.28 \text{ mm day}^{-1}$. In contrast, the average precipitation from AMIP in the same region is 5.17 mm day^{-1} , and the corresponding contribution from ocean surface evaporation is 3.90 mm day^{-1} . The contribution from moisture flux convergence below 850 hPa is only 2.23 mm day^{-1} . This clearly demonstrates that the enhanced moisture convergence due to coupled feedback is the major contributor to the enhanced southern ITCZ precipitation. This explains why precipitation is increased significantly in the CPL simulation in the southern ITCZ region even when SST is decreased relative to the AMIP simulation.

The question is then “What caused the overly cold SST in the northern ITCZ region in boreal spring?” To quantitatively address this question, we analyze the ocean mixed layer heat budget in the northern ITCZ region of eastern Pacific (5° – 15°N , 90° – 120°W). To ensure that the ocean mixed layer temperature from GODAS is a good proxy for SST from HadISST, we examine the seasonal cycle of SST from HadISST and mean temperature in ocean mixed layer from GODAS averaged in the region where the heat budget is calculated. It is found that the correlation between them is 0.99 with root-mean-square difference 0.14°C and annual mean difference -0.05°C , indicating that the ocean mixed layer temperature from GODAS can reasonably represent the HadISST SST in the analyzed region.

The equation for the mixed layer temperature can be written as (Moisan and Niiler 1998)

$$\frac{\partial T_a}{\partial t} = \frac{Q - Q_s(-h)}{\rho c_p h} - u_a \frac{\partial T_a}{\partial x} - v_a \frac{\partial T_a}{\partial y} - \frac{(T_a - T_{-h})}{h} (w_e + w_{-h}) + R, \quad (1)$$

where T_a is the vertically averaged mixed layer temperature (a proxy of SST), h is the mixed layer depth, $w_e = dh/dt = \partial h/\partial t + u_{-h}(\partial h/\partial x) + v_{-h}(\partial h/\partial y)$ is the entrainment velocity, w_{-h} , u_{-h} , and v_{-h} are the vertical, zonal, and meridional current velocities at the mixed layer base, respectively, T_{-h} is the temperature just below the mixed layer, ρ is the reference density of seawater, and c_p is the specific heat of seawater at constant pressure. Here Q represents the net surface energy flux, which is positive into the ocean, and $Q_s(-h)$ is

shortwave radiative flux penetrating through the base of the mixed layer. Also, u_a and v_a are the vertically averaged velocities of zonal and meridional currents, respectively; R represents the residual term, which includes horizontal diffusion, the integral of the vertical temperature–velocity covariance, and submonthly transient eddies. Following Wang and McPhaden (1999), the shortwave radiative flux penetrating through the base of the mixed layer is parameterized as $Q_s(-h) = 0.45Q_s(\text{sfc}) \exp(-0.04h)$, where $Q_s(\text{sfc})$ is the net surface shortwave radiative flux.

Liu et al. (2010) used four ocean reanalysis datasets including GODAS to calculate the mixed layer heat budget in the double-ITCZ region. They found that the results have only small differences between the calculations using daily and monthly mean data. Liu et al. (2012) used the model data from the NCAR CCSM3.0, a predecessor of NCAR CESM model, to calculate the mixed layer budget in the double-ITCZ region and also found only small differences between the calculations with daily and monthly mean model output. Therefore, monthly mean data from NCEP GODAS and model simulations are used to perform ocean mixed layer heat budget analysis in this study. We examined the annual mean heat budget and found that the residual term is negligible compared to entrainment term and surface flux term, indicating that the budget analysis is reasonably accurate. Therefore, only contributions from the first four terms on the right-hand side of Eq. (1)—the net energy flux into the mixed layer, zonal advection, meridional advection, and vertical entrainment—are presented in this study.

Figure 5 shows the seasonal variation of temperature tendencies contributed by each term to the mixed layer temperature in the northern ITCZ region for NCEP GODAS and the CPL simulation. Note that the temperature decreases when the temperature tendency is negative and reaches its minimum when tendency is increased to zero from negative. The temperature increases when temperature tendency is positive and reaches its maximum when tendency is reduced to zero. The GODAS shows that the total temperature tendency is increased from negative to zero between January and February, indicating that the SST reaches its minimum between January and February. Actually, Fig. 1a shows that the HadISST in the northern ITCZ region reaches its minimum (about 27°C) around January. The positive temperature tendency then leads to a gradual increase in SST until June, when the tendency is reduced to zero and SST reaches its maximum around 29°C. After June, the small negative temperature tendency leads the SST to slowly decrease to its minimum in January. The seasonal cycle of SST in the northern

FIG. 5. Seasonal cycles of mixed layer temperature tendencies averaged over 5°–15°N, 90°–120°W contributed from each term on the right-hand side of Eq. (1) for the (a) GODAS and (b) CPL simulation.

ITCZ region from the GODAS is generally determined by the variation of surface heat flux, especially from February to September. The subsurface vertical process dominates the negative tendency from October to January. The total temperature tendency from the CPL simulation is increased from negative to zero in February, corresponding to the SST minimum of about 25°C. The total temperature tendency from the CPL is about 0.65° and 0.8°C month^{−1} for March and April, respectively. In contrast, the total temperature tendency from the GODAS is about 0.55° and 0.7°C month^{−1} for March and April, respectively. The stronger warming rate from the CPL relative to the GODAS during boreal spring is not able to explain the severe cold bias of SST in the CPL relative to the HadISST observation, indicating that the cold bias is inherited from boreal winter. Figure 1 shows that the decrease of SST in northern ITCZ region from CPL is more rapid than the HadISST from October to January. Correspondingly, the CPL produces much stronger negative tendency of temperature than the GODAS from October to January. The total temperature tendencies produced by CPL for October–January are about −0.4°, −0.95°, −0.7°,

and $-0.3^{\circ}\text{C month}^{-1}$, respectively, while the tendencies produced by GODAS are only -0.1° , -0.05° , -0.2° , and $-0.1^{\circ}\text{C month}^{-1}$, respectively. The stronger cooling rate produced by the CPL simulation during boreal winter results in the pronounced cold bias of SST relative to the observations. Among the four terms, the vertical entrainment and surface heat flux contribute to the negative temperature tendency from October to January, with slightly larger contribution from the vertical entrainment.

The wind stress can force Ekman currents. The positive wind stress curl in the Northern Hemisphere corresponds to the divergence of Ekman transport and hence Ekman upwelling. The upwelling of cold water can lead to negative tendency of mixed layer temperature. Figure 6 shows the distribution of wind stress vector and curl from the GODAS, CPL, and AMIP simulations for November, when the cooling difference between CPL and GODAS is most significant. The corresponding precipitation distributions from the GPCP, CPL, and AMIP simulations are also presented to demonstrate the relationship between convection and ocean cooling biases. The GPCP shows a southeastward precipitation band near 10°N . Correspondingly, the surface wind of GODAS converges to the precipitation band and forms a southeastward positive band of wind stress curl, indicating an Ekman upwelling band. The CPL produces a much stronger precipitation band near 10°N relative to the GPCP, leading to wind stress curl almost doubled compared to the GODAS, resulting in much stronger mixed layer cooling. This explains why the cooling rate of vertical entrainment process from the CPL is almost doubled compared to the GODAS in November (Fig. 5). To examine to what extent the precipitation and wind stress errors can be traced back to the atmospheric models, Figs. 6e and 6f show the distribution of precipitation and wind stress curl from the AMIP simulation. They show that the AMIP simulation also produces much stronger precipitation band compared to GPCP and even stronger positive wind stress curl than the CPL. This indicates that the pronounced cooling bias from vertical entrainment process in the CPL originates from atmospheric model errors in precipitation and wind stress, which all are associated with the model convection bias. The coupled processes tend to enhance the precipitation bias and reduce the wind stress bias to some extent.

Another contributor to the cooling bias is the net surface heat flux. The surface heat flux consists of shortwave radiative flux, longwave radiative heat flux, sensible heat flux, and latent heat flux. Figure 7a presents the seasonal cycle of the difference of each surface heat flux

component between the CPL and NCEP averaged in the northern ITCZ region. It shows that the cooling bias in the CPL related to surface heat flux in boreal winter is largely determined by the increase of upward latent heat flux, which contributes about -25 W m^{-2} to the cooling bias, while the contribution from sensible heat flux to the cooling bias of mixed layer is only about -6 W m^{-2} . In the AMIP simulation (Fig. 7b), the latent heat flux is still the largest contributor to the net negative heat flux in boreal winter, which is even larger than that in the CPL for January–March (Fig. 7c). This indicates that the negative latent heat flux bias in the CPL in the analyzed region originates from the atmospheric model and the coupled interaction tends to mitigate this bias. The sensible heat flux bias in the AMIP simulation is very similar to that of the CPL, suggesting that the negative sensible heat flux bias in the CPL also originated from the atmospheric model. Although the longwave radiative flux in the CPL has no contribution to the cooling bias in October and November, it contributes about -10 W m^{-2} to the cooling bias from January to March. In contrast, the AMIP simulation produces positive bias of longwave radiative flux year-round. This indicates that the negative bias of longwave radiative flux during January–March in the CPL is a result of coupled interaction. This negative bias of longwave radiative flux in the CPL may be attributed to the precipitation decreases relative to the AMIP simulation during January–March, which indicates the reduced cloud fraction and increased loss of longwave heat flux. Note that the seasonal cycle of the net surface heat flux bias in the CPL (Fig. 7a) show slight difference compared to that of corresponding temperature tendency bias. For example, during January and February, the CPL shows positive bias of net surface heat flux while the negative bias of temperature tendency related to the surface heat flux. This difference can be largely attributed to the deepening of mixed layer depth in the CPL because the temperature tendency due to surface heat flux is inversely proportional to the mixed layer height [Eq. (1)].

Because latent heat flux is the major contributor to the cooling bias related to surface heat flux and is determined by the near-surface wind speed and humidity gradient, we further examine the biases of the latent heat flux, wind speed at 10-m height, and the humidity gradient between specific humidity at 2-m height and saturation specific humidity at the sea surface for November, when the latent heat flux bias is the most significant. All three biases averaged over October, November, and December are also examined, which show very similar distributions to those of November, although slight differences exist in magnitude, indicating

FIG. 6. (left) Wind stress vector and curl from (a) GODAS, (c) CPL, and (e) AMIP and (right) precipitation from (b) GPCP, (d) CPL, and (f) AMIP for November.

FIG. 7. Seasonal cycles of differences of surface heat flux components averaged over 5°–15°N, 90°–120°W between (a) CPL and NCEP, (b) AMIP and NCEP, and (c) CPL and AMIP.

that the analyses of November may explain the winter latent heat flux bias well. Figure 8 shows the biases of latent heat flux, wind speed at 10 m, and humidity gradient produced by the CPL and AMIP simulations relative to the NCEP reanalysis for November. The latent heat flux bias of the CPL in the northern ITCZ region is generally negative, which indicates more latent heat loss from the mixed layer. The latent heat flux bias of CPL is relatively small in the center zone of precipitation band and much larger on both sides of the center zone, with largest bias off the western coast of Central America. The distribution of latent heat flux bias generally resembles the distribution of surface wind speed

bias, suggesting that the latent heat flux bias is mainly determined by the wind speed bias. Note that the precipitation band is usually accompanied by a convergence band of near-surface wind. The wind speed is generally small in the center zone of the convergence band and much larger on both sides of the center zone. When the convection (precipitation) is strengthened, the convergence and wind speed will be enhanced as well. The near-surface wind speed of the CPL actually shows a relatively smaller positive bias in the center zone of precipitation band and larger positive bias on both sides of the center zone, corresponding to the increase of precipitation (Fig. 6). The largest wind speed bias occurs off the western coast of Central America between 13° and 15°N, which can be related to the convection bias over land of Central America in the succeeding analysis (Fig. 9). The increase of wind speed will enhance the surface evaporation and increase the upward surface latent heat flux. In addition, deep convection tends to dry the lower troposphere and increase the air-sea humidity difference. The strengthened convection may further increase the humidity gradient. Actually Fig. 8e shows that the humidity gradient is enhanced in the CPL compared to the NCEP reanalysis. This result suggests that the increase of surface latent heat flux in the CPL simulation can be largely attributed to the enhanced near-surface wind speed, which can be generally related to the convection bias.

The biases of latent heat flux produced by the AMIP simulation relative to the NCEP reanalysis shows different distribution to that of the CPL. In the center zone of precipitation band, the AMIP simulation produces positive latent heat flux bias, while negative bias exists in the CPL, indicating that the negative latent heat flux bias in this region in the CPL originates from the coupled interaction. On both sides of the center zone of the precipitation band, the AMIP produces negative latent heat flux bias, with stronger bias in the northeast and weaker bias in the south relative to the CPL, indicating the negative latent heat flux bias in these regions in the CPL originates from the atmospheric model bias and the coupled interaction may modulate it. The distribution of latent heat flux bias produced by AMIP simulation is also similar to that of wind speed bias, suggesting that the wind speed bias may be the major contributor of latent heat flux bias in the atmospheric model as well. Similar to the CPL, the AMIP simulation also produces the largest positive wind speed bias off the western coast of Central America between 13° and 15°N, indicating that this bias originates from the atmospheric model. To investigate the formation mechanism of this wind speed bias, Fig. 9 shows the zonal-vertical distribution of vertical velocity, zonal and meridional winds

FIG. 8. (left) The difference between CPL and NCEP in (a) latent heat flux, (c) wind speed at 10 m, and (e) humidity gradient and (right) the difference between AMIP and NCEP in (b) latent heat flux, (d) wind speed at 10 m, and (f) humidity gradient for November. The humidity gradient is calculated as the difference between the specific humidity at 2-m height and saturation specific humidity at sea surface.

FIG. 9. Zonal-vertical distribution of (a) vertical velocity (shading; $0.25 \text{ hPa day}^{-1}$) and wind vectors (m s^{-1}), (b) zonal wind (shading; m s^{-1}) and wind vectors (m s^{-1}), and (c) meridional wind (m s^{-1}) averaged between 13° and 15°N for November from ERA-Interim. (d)–(f) The corresponding biases from the AMIP simulation.

averaged between 13° and 15°N for November from ERA-Interim, and the biases from the AMIP simulation. The distribution of biases from CPL is similar to that of AMIP and is not shown here. ERA-interim shows that the Central American topography blocks the strong easterly wind from the Caribbean Sea and forces it to climb up near the Atlantic coastal area (Fig. 9b). The topographic friction reduces the near-surface wind speed and easterly winds flow over the Central American terrain with high speeds near 750 hPa. In November, the descending motion prevails over the middle and eastern portions of Central America (Fig. 9a). The divergence of descending flow near the surface enhances the northerly winds (Fig. 9c). The AMIP simulation produces remarkable anomalous ascending motion over the middle portion of Central America (85° – 90°W) with maximum upward velocity between 900 and 850 hPa (Fig. 9d), which can be attributed to model convection bias that will be demonstrated in ensuing analysis with Fig. 10. Where the convective vertical velocity decreases with height, there is mass divergence and convective outflow. Figure 9e shows that the outflow of anomalous shallow convection enhances the easterly wind in downwind side of convection in the lower troposphere. The westward convective outflow may also strengthen the downhill flow in the eastern portion of Central America (Fig. 9d). The enhanced near-surface divergence associated with the strengthened descending flow may increase the near-surface easterly wind (Fig. 9e)

and northerly wind (Fig. 9f), contributing to the increase of surface wind speed off the western coast of Central America. The enhanced descending flow may also carry more enhanced easterly momentum from lower troposphere to the surface, leading to the increase of surface wind speed. Therefore, the large surface wind speed bias off the western coast of Central America can be largely related to the upstream convection bias over land of Central America. In addition, the zonal circulation changes with anomalous convergence and

FIG. 10. Surface total precipitation rate from GPCP (blue), ERA-Interim (black), and the AMIP simulation (red), and convective precipitation rate from the AMIP simulation (purple) averaged between 13° and 15°N for November.

ascending motion near 100°–110°W and anomalous subsidence near 70°–80°W may also increase the background easterly wind in lower troposphere and contribute the increase of surface wind speed. The enhanced subsidence between 92.5° and 100°W may also contribute to the drying of the near-surface air and the increase of air–sea humidity difference between 13° and 15°N (Fig. 8f).

To demonstrate that the anomalous ascending motion over the middle portion of Central America (85°–90°W) is produced by convection rather than forced by model gridscale topography, Fig. 10 presents the surface precipitation rate from GPCP, ERA-Interim, and AMIP simulations averaged between 13° and 15°N for November. Since precipitation in the AMIP simulation consists of convective precipitation associated with convection and gridscale precipitation associated with gridscale ascending motion, the convective precipitation rate from AMIP simulation is also presented. It is clear that the precipitation in AMIP simulation between 85° and 90°W comes almost entirely from convection, and is much higher than the precipitation of GPCP and ERA-Interim. Corresponding to the largest bias of ascending motion near about 87.5°W, the convective precipitation produced by AMIP simulation is about 5 mm day^{-1} , which overestimates the precipitation in GPCP and ERA-Interim (3.25 mm day^{-1}) by more than 50%. This demonstrates that the anomalous ascending motion over the middle portion of Central America (85°–90°W) in the AMIP simulation can be attributed to the overestimated convection.

We also examined the bias distribution of surface wind speed and latent heat flux in other months, and found that a similar bias off the western coast of Central America between 13° and 15°N persists in both AMIP and CPL simulations from November to February, when the winter northeasterly trade wind prevails. During January and February, the CPL produces much weaker oceanic convection than the AMIP simulation in the northern ITCZ region (Fig. 2). However, both CPL and AMIP still produce the largest positive wind speed bias and negative latent heat flux bias off the western coast of Central America between 13° and 15°N, which can be related to the land convection bias over Central America. Since the large latent heat flux bias off the western coast of Central America contributes significantly to the area-averaged bias over 5°–15°N, 90°–120°W, both CPL and AMIP show excessive area-averaged latent heat flux in January and February in Fig. 7.

4. Summary and conclusions

GCMs forced with observed SST commonly exhibit a very weak double-ITCZ bias, which is severely

exacerbated in corresponding coupled GCMs, indicating that air–sea interaction and simulated SST bias play important roles in the formation of double-ITCZ bias. The eastern Pacific double-ITCZ bias has long been attributed to the warm bias of SST in the southeastern Pacific (SEP) and associated local air–sea interaction. In this study, we conducted a set of experiments using the NCAR CESM1.2.1 to investigate whether the double-ITCZ bias is still significantly exacerbated in the eastern Pacific when the air–sea coupled feedback produces cold SST bias in the SEP. We first conducted an AMIP simulation in which the active ocean and sea ice models in the CESM1.2.1 were replaced by observed SST and sea ice forcing from HadISST dataset in 1980s. Then a fully coupled simulation of CESM1.2.1 was performed with preindustrial forcing at 1850 level. It is shown that the simulated preindustrial SST in the SEP is slightly colder than the observed SST in the 1980s. Note that the very similar double-ITCZ bias exists in both coupled preindustrial and twentieth-century simulations of CESM1.2.1 (Song and Zhang 2016), indicating that the greenhouse gas level has very little impact on the double-ITCZ bias in coupled GCMs. Thus, the major difference between the AMIP and coupled simulations comes from the air–sea interaction and associated SST changes. The ITCZ precipitation simulated by AMIP simulation is systematically enhanced compared to the GPCP observations, resulting in a stronger double ITCZ in boreal spring. However, in the coupled simulation, the air–sea interaction dramatically influences the precipitation distribution. The northern ITCZ precipitation disappears from January to May and the southern ITCZ precipitation is significantly increased from January to June, resulting in a pronounced seasonally alternating type (Machoso et al. 1995) of double-ITCZ bias. This result shows that the significant double-ITCZ bias can still form in the eastern Pacific through air–sea coupled feedback even when there is a cold SST bias in the SEP, indicating that this double-ITCZ bias in the eastern Pacific should not be caused by changes in local SST forcing in the SEP; other nonlocal culprits and mechanisms should be responsible for the double-ITCZ bias in the eastern Pacific.

Further analyses show that the coupled simulation produces an overly cold bias of SST relative to the AMIP simulation in the northern ITCZ region in boreal spring. The simulated SST is even colder than 26°C from January to March. Since deep convection is rarely observed where SST is colder than 26°C in the tropics, this overly cold bias of SST may suppress deep convection in the northern ITCZ region. The enhanced low-level cross-equatorial northerly wind strengthens the wind

convergence south of the equator and transports abundant water vapor to the convergence zone, leading to significant increase of precipitation in the southern ITCZ region even though the underlying SST has a slight cold bias. This mechanism is not inconsistent with previous research. The ITCZ tends to develop over the warmest off-equatorial surface water (Xie and Philander 1994, Waliser and Somerville 1994). Since the physical feedback and ocean–atmosphere interactions favor neither hemisphere with the SST maximum and the ITCZ (Xie and Philander 1994, Waliser and Somerville 1994), the asymmetric forcing between hemispheres associated with the geography of continents plays an important role in the formation of the single ITCZ and SST maximum that are located in the Northern Hemisphere in the eastern Pacific in the observations (Philander et al. 1996, Takahashi and Battisti 2007a). However, the coupled model simulation produces a severe cold SST bias in the northern ITCZ region in boreal spring so that convection is suppressed there and the SST maximum is located in the Southern Hemisphere rather than the Northern Hemisphere in the eastern tropical Pacific. It is therefore not surprising that a single ITCZ develops over the warmest water in the Southern Hemisphere.

The key question is then “What caused the overly cold SST in the northern ITCZ region in boreal spring?” To address this question, the ocean mixed layer heat budget in the northern ITCZ region of eastern Pacific is analyzed. The coupled simulation shows stronger warming rate during boreal spring relative to GODAS, which is not able to explain the severe cold bias of SST in the simulation relative to the observation, indicating that the cold bias is inherited from boreal winter. It is shown that the SST in the northern ITCZ region decreases more rapidly in the coupled simulation than in HadISST from October to January. Correspondingly, the coupled simulation produces a much stronger negative tendency of mixed layer temperature than GODAS from October to January. The vertical entrainment processes and surface heat flux are the two contributors to the negative bias of mixed layer temperature tendency, with slightly larger contribution from the vertical entrainment process.

Because ocean Ekman upwelling of cold water can lead to a negative tendency of mixed layer temperature and is related to the wind stress curl, we further examined the biases of wind stress and precipitation in the northern ITCZ region. It is shown that the coupled simulation produces a much stronger precipitation band in the northern ITCZ region relative to the GPCP in November. The corresponding change in surface wind stress leads to wind stress curl is almost doubled

compared to the GODAS, resulting in much stronger Ekman upwelling and mixed layer cooling. To examine to what extent the precipitation and wind stress errors can be traced back to the atmospheric model, the distribution of precipitation and wind stress curl from the AMIP simulation is also examined. It shows that the AMIP simulation also produces a much stronger precipitation band compared to GPCP and even stronger positive wind stress curl than the coupled simulation. This indicates that the pronounced cooling bias from the vertical entrainment process in the coupled simulation originates from atmospheric model errors in precipitation and wind stress, which all are associated with the model convection bias.

The analyses of surface heat flux biases show that the cooling bias in the coupled simulation related to surface heat flux in boreal winter is largely determined by the increase of upward latent heat flux, with a relatively smaller contribution from sensible heat flux. Comparing to the surface heat flux biases in the AMIP simulation, it is shown that the latent and sensible heat flux biases in the coupled simulation generally originate from the atmospheric model except in the center zone of precipitation band where the weak negative latent heat flux bias in the coupled simulation originates from the coupled interaction. The enhanced surface latent heat flux can be attributed to the increases in the near-surface wind speed and humidity gradient. The increase of wind speed can be partly related to the upstream convection bias over land of Central America and the enhanced wind convergence associated with the oversimulated convection in the northern ITCZ region. The increase of humidity gradient can also be partly attributed to the enhanced drying of lower troposphere due to the oversimulated convection. Therefore, the cooling bias from the surface latent heat flux in the coupled simulation can be generally related to the model convection bias as well.

In summary, the oversimulated convection in the northern ITCZ region and Central America in boreal winter may result in biases in the surface wind fields in tropical northeastern Pacific in the atmospheric model, which favor the cooling of ocean mixed layer through the enhancement of latent heat flux and Ekman upwelling. Once the atmospheric model is coupled to the ocean model, these biases are passed into the ocean model and result in a severe cold bias of SST in the northern ITCZ region. The overly cold SST bias persists in the subsequent spring, leading to the suppression of convection in the northern ITCZ region. The enhanced low-level cross-equatorial northerly wind strengthens the wind convergence south of the equator and transports abundant water vapor to the convergence zone, strengthening the southern ITCZ convection. All these

processes lead to the disappearance of northern ITCZ and the enhancement of southern ITCZ in boreal spring, forming a seasonally alternating type of double-ITCZ bias. This study suggests that convection biases in the northern ITCZ and Central America in boreal winter may be a culprit for the double-ITCZ bias in the eastern Pacific. Recently Song and Zhang (2018) improved the Zhang and McFarlane (1995) convection scheme in the CESM1.2.1 and found that the double-ITCZ bias was largely eliminated in all seasons in the tropical Pacific. The formation mechanism of the eastern Pacific ITCZ bias proposed in this study will be further examined with the improved convection scheme in a future paper. It should be noted that the proposed formation mechanism for the eastern Pacific double-ITCZ bias in this study is based only on one GCM, the NCAR CESM1.2.1. Since 11 of 17 GCMs that participated in CMIP5 (phase 5 of the Coupled Model Intercomparison Project) show a similar seasonally alternating type of double-ITCZ bias in the eastern Pacific (Oueslati and Bellon 2015), examining the proposed mechanism in other GCMs is highly desired in future studies.

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