Convection Parameterization, Tropical Pacific Double ITCZ, and Upper-Ocean Biases in the NCAR CCSM3. Part I: Climatology and Atmospheric Feedback

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(Manuscript received 21 May 2008, in final form 3 February 2009)

ABSTRACT

The role of convection parameterization in the formation of double ITCZ and associated upper-ocean biases in the NCAR Community Climate System Model, version 3 (CCSM3) is investigated by comparing the simulations using the original and revised Zhang–McFarlane (ZM) convection schemes. Ten-year model climatologies show that the simulation with the original ZM scheme produces a typical double ITCZ bias, whereas all biases related to the spurious double ITCZ and overly strong cold tongue in precipitation, sea surface temperature (SST), wind stress, ocean thermocline, upper-ocean currents, temperature, and salinity are dramatically reduced when the revised ZM scheme is used. These results demonstrate that convection parameterization plays a critical role in the formation of double ITCZ bias in the CCSM3. To understand the physical mechanisms through which the modifications of the convection scheme in the atmospheric model alleviate the double ITCZ bias in the CCSM3, the authors investigate the impacts of convection schemes on the atmospheric forcing and feedback in the uncoupled Community Atmospheric Model, version 3 (CAM3). It is shown that the CAM3 simulation with the original ZM scheme also produces a signature of double ITCZ bias in precipitation, whereas the simulation with the revised ZM scheme does not. Diagnostic analyses have identified three factors on the atmospheric side (i.e., the sensitivity of convection to SST, the convection–shortwave flux–SST feedback, and the convection–wind–evaporation–SST feedback) that may contribute to the differences in the coupled simulations.

1. Introduction

The intertropical convergence zone–South Pacific convergence zone (ITCZ–SPCZ) complex is one of the most striking characteristics of tropical climate. In observations, the ITCZ is manifested as a zonal band of intense convection located north of the equator over the warmest sea surface temperature (SST) for much of the year and accompanied by a band of surface wind convergence. The SPCZ is a semipermanent convection band extending from the ITCZ near the equator, around eastern Papua New Guinea, southeastward to the extratropical South Pacific (around 30°S, 130°W). Thus, a prominent feature of the tropical Pacific climate is the asymmetry between the ITCZ and SPCZ around the equator. However, since the early days of ocean–atmosphere coupled general circulation model (CGCM) development, most CGCMs simulate two parallel bands of intense rainfall (double ITCZs) straddling the equator across much of the Pacific, with an excessive westward extension of a band of relatively weak rainfall (dry tongue) over the equator that splits the ITCZ–SPCZ complex in the western Pacific (Mechoso et al. 1995; Meehl and Arblaster 1998; Kirtman et al. 2002; Kiehl and Gent 2004; Meehl et al. 2005; Collins et al. 2006; Zhang et al. 2007). Usually this so-called double ITCZ bias in precipitation is also accompanied by significant biases in the upper ocean: a warm SST bias in a band near 10°S in the central and eastern Pacific, excessive westward extension of the cold SST tongue, and spurious symmetric structures of the thermocline and upper-ocean currents (Mechoso et al. 1995; Zhang et al. 2007).

Because the mean climate provides a background state for climate variability, the double ITCZ bias in the climate mean state was believed to be an important factor affecting the simulation of climate variability in CGCMs. For example, associated with the double ITCZ bias in the mean state, the CGCMs tend to split the intraseasonal convective anomaly into a double ITCZ...
structure in the Pacific and simulate excessively strong convective signals in the eastern Pacific ITCZ (Inness and Slingo 2003), or it fails to capture the observed equatorial maximum of MJO variance in the western Pacific (Lin et al. 2006). On interannual time scales, the CGCMs with double ITCZ biases tend to simulate too-frequent “aborted” El Niño–Southern Oscillations (ENSOs; Guilyardi et al. 2003) and overestimate the ENSO variability in the central and western Pacific (Kiehl and Gent 2004). Therefore, the double ITCZ deficiency may limit the ability of CGCMs to simulate and predict climate change.

Over the last decade, despite the considerable effort that has gone into improving the model physics and resolutions, significant double ITCZ bias in the tropics remains a serious problem. Recent evaluations of the twentieth-century climate simulations by the newest generation of CGCMs participating in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4; Dai 2006; Lin 2007) showed that most of the current state-of-the-art CGCMs, including the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3 (CCSM3), still produce double ITCZs. Alleviating the double ITCZ bias in CGCMs remains an enormous challenge for coupled global climate modeling.

The cause and mechanism for the formation of double ITCZs in CGCMs are not yet well understood. Several studies (Ma et al. 1996; Dai et al. 2003, 2005) suggested that the spurious southern ITCZ rainband could be attributed to the warm bias of SST in the southeast Pacific, resulting from underestimated stratus cloud cover off the coast of Peru. They consider the warm bias of SST in a band near 10°S in the central and eastern Pacific as a westward extension of the warm bias in SST off the South American coast. Ma et al. (1996) artificially increased the amount of stratus clouds off the coast of Peru in a CGCM and found that the warm bias in SST in the southeastern Pacific was significantly alleviated, leading to a more realistic SPCZ. However, the extensive bias of cold tongue was exacerbated in their simulation. Dai et al. (2005) obtained similar results by improving the simulation of low-level cloud fractions and net surface shortwave (SW) radiation fluxes in the subtropical eastern Pacific using a statistically based low-level cloud parameterization scheme in a CGCM. They found that the SST bias and therefore the double ITCZ precipitation bias in the southeastern Pacific are reduced; however, the equatorial cold tongue is strengthened and extends farther westward, and the precipitation is weakened in the western equatorial Pacific. These results suggest that improving stratus cloud cover alone is not enough to eliminate the double ITCZ bias.

Convection parameterization is another factor believed to contribute to the formation of the double ITCZ bias in precipitation. Several studies have attempted to eliminate this bias in atmospheric models by modifying the convection scheme. Bacmeister et al. (2006) showed that increasing rain reevaporation in convection parameterization may reduce the double ITCZ bias in version 2 of the National Aeronautics and Space Administration’s (NASA) Seasonal-to-Interannual Prediction Project (NSIPP-2) atmospheric GCM. Zhang and Mu (2005) implemented the revised Zhang–McFarlane (ZM) scheme (Zhang 2002), in which the most important modification is the use of a new closure derived from observational studies in the NCAR Community Climate Model, version 3 (CCM3), and the results showed that the double ITCZ bias in the atmosphere-only model was significantly reduced. Zhang and Wang (2006) further tested the scheme in the NCAR coupled model CCSM3 and showed that both precipitation and SST biases in the central Pacific south of the equator were reduced in boreal summer. This study extends the work of Zhang and Wang (2006) to systematically examine the impacts of convection schemes on the simulation of the ITCZ–SPCZ complex and the associated ocean state in the NCAR CCSM3 by comparing the simulations with the original ZM convection scheme (Zhang and McFarlane 1995) to the simulation with the revised ZM scheme. An attempt is also made to explore the possible interactions and feedback mechanisms among convection, large-scale circulation, and SST that may help explain the CCSM3 simulation differences.

The paper is organized as follows: Section 2 briefly describes the model, the convection schemes, the simulations, and the observational data used to compare with model results. Section 3 presents the impacts of convection schemes on the double ITCZ bias in the CCSM3. Section 4 examines the impacts of convection schemes on the atmospheric forcing and potential feedback to the ocean related to the double ITCZ bias. Section 5 presents a summary of results and conclusions.

2. Models, convection schemes, and data

The NCAR CCSM3 is a state-of-the-art coupled global climate model consisting of four component models (atmosphere, ocean, land, and sea ice) linked by a central coupler. The atmosphere model, the Community Atmosphere Model, version 3 (CAM3), is a global atmospheric general circulation model with T42 truncation (approximately $2.8^\circ \times 2.8^\circ$ latitude–longitude) in the horizontal and 26 levels in the vertical. Deep convection in CAM3 is parameterized using the Zhang–McFarlane scheme (Zhang and McFarlane 1995). The
land surface model is the Community Land Model (CLM), which uses the same horizontal grids as the atmospheric model. The ocean model is an extension of the Parallel Ocean Program (POP) version 1.4.3 from the Los Alamos National Laboratory (LANL), which has 40 vertical levels and a longitudinal resolution of approximately 1°. The latitudinal resolution of POP is variable, with finer resolution (approximately 0.3°) near the equator. The sea ice model, the Community Sea Ice Model (CSIM), shares a common horizontal grid with the ocean model (for details of the model, see Collins et al. 2006).

The ZM convection scheme uses convective available potential energy (CAPE)-based closure, which assumes that convection consumes the CAPE with a relaxation time of two hours. Thus, the occurrence and intensity of convection are determined by the amount of CAPE. However, recent observational studies (Zhang 2002, 2003) showed that this assumption is not a good approximation for time scales shorter than a day. These studies also found that convection is well correlated with the change of CAPE associated with large-scale forcing in the free troposphere. Based on this finding, Zhang (2002) proposed a revised Zhang–McFarlane scheme, in which the major modification is the use of the new closure derived from the observational data. The new closure assumes that stabilization of the atmosphere by convection is in quasi equilibrium with the destabilization by the large-scale forcing in the free troposphere. Thus, the occurrences and intensities of convection are determined by positive forcing on CAPE from large-scale processes in the free troposphere. More details on the revised ZM scheme can be found in Zhang (2002). It should be noted that although the revised ZM scheme produces reasonable climatology and intraseasonal variability in the NCAR global models, it fails to reproduce the observed diurnal cycle of convection over land areas such as the North American monsoon region, where surface thermodynamic forcing is important for convection (Collier and Zhang 2006). Because this paper focuses only on the impact of convection schemes on the mean climate over tropical Pacific, this defect should have no significant impact on the conclusions of this paper.

To investigate the impacts of convection schemes on the double ITCZ bias, two 12-yr simulations using the CCSM3 are conducted. In the control simulation (referred to as the CTL run) the standard CCSM3 configuration (i.e., the Zhang–McFarlane convection scheme) is used, whereas in the experimental simulation (referred to as the RZM run) the revised Zhang–McFarlane scheme is used. Both simulations start from 1 January and run for 12 yr, with the initial ocean temperature and salinity conditions taken from the January climatology of Levitus et al. (1998). The 10-yr averages for years 3–12 are used to examine the double ITCZ bias. We also conducted two 16-yr-long CAM3 simulations from September 1979 to August 1995, one with the ZM scheme (referred to as CAM3z) and the other with the revised ZM scheme (referred to as CAM3z), using observed SST as the boundary conditions. The monthly data from January 1980 to December 1994 of these two simulations will be used to explore the possible atmospheric feedback on the ocean.

To identify the double ITCZ bias in model simulations, this study uses the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) from 1979 to 1998, the Global Precipitation Climatology Project (GPCP) version 2 combined precipitation dataset from 1979 to 2003 (Adler et al. 2003), Reynolds et al.’s (2002) optimum interpolation SST (OISST) analysis from 1971 to 2000, and the ocean temperature and current velocity data of the National Centers for Environmental Protection (NCEP) Global Ocean Data Assimilation System (GODAS; Ji et al. 1995; Behringer et al. 1998) from 1980 to 1999. Also used to evaluate the model results are the surface wind stress data of the NCEP Ocean Data Assimilation System (ODAS) tropical Pacific Ocean monthly analyses (Ji et al. 1995; Derber and Rosati 1989) from 1980 to 1999, the salinity of Levitus et al. (1998) from 1900 to 1997, and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Källberg et al. 2004) from 1980 to 2001.

3. Impacts of convection schemes on double ITCZ bias

Figure 1 shows the annual-mean precipitations from the CMAP and GPCP observations and the CTL and RZM simulations, and it also shows the differences between CTL and CMAP and between RZM and CTL. Two sets of observational estimates are presented to illustrate the uncertainty in observations. The CTL produces a typical double ITCZ bias in the tropical Pacific. Its simulated precipitation (Fig. 1c) shows two remarkably zonal ITCZ rainbands near 7° south and north of the equator. A band of precipitation of >9 mm day−1 extends as far east as 135°W south of the equator, with a maximum near 155°W. In CMAP observations (Fig. 1a), only a single zonal ITCZ rainband appears north of the equator, with a southeast-oriented SPCZ extending from east of Papua New Guinea to the extratropical South Pacific (around 30°S, 130°W). The region with precipitation of >9 mm day−1 is located west of the
international date line in the Southern Hemisphere. In addition, the dry tongue in precipitation over the equator in the CTL is much stronger and displaced approximately 10° westward, relative to the observations. Compared to the CTL, the precipitation simulated in the RZM (Fig. 1d) shows a significant increase over 2°S–2°N and a decrease east of the international date line over the southern ITCZ region, with a precipitation maximum of 9 mm day⁻¹ located west of 170°W, and is more comparable to the observations. The double ITCZ bias is reduced considerably. Note that the precipitation over northern ITCZ in the RZM is relatively weaker than in the CTL and the CMAP observation. But compared to the GPCP observations (Fig. 1b), it may not be too much of degradation. Outside the Pacific, the most noticeable degradation is in the eastern Indian Ocean just west of Sumatra, where the dry bias already existing in the CTL increases in the RZM. However, the location of the Indian Ocean ITCZ is better simulated in the RZM (south of the equator) than in the CTL (over the equator).

The best way to quantify the double ITCZ bias in the CCSM3 model simulation and its improvement using the revised ZM scheme is to present the precipitation differences between the CTL and the CMAP and between the RZM and the CTL (Figs. 1e,f), respectively. Corresponding to the spurious southern ITCZ in the CTL, Fig. 1e shows positive rainfall biases of up to 9 mm day⁻¹ between 3° and 12°S east of the international date line and negative biases of up to 3 mm day⁻¹ over the SPCZ, relative to the observations, indicating that the southeast-oriented SPCZ is replaced by a zonally oriented southern ITCZ. The precipitation produced by the RZM is reduced from the CTL by up to 6 mm day⁻¹ between 3° and 12°S and increased by up to 2 mm day⁻¹ over the SPCZ (Fig. 1f). This represents an approximate two-third reduction in precipitation biases in these regions and an improvement in the simulation of SPCZ. Associated with the westward extension of the overly dry tongue, the CTL (Fig. 1e) shows negative rainfall biases of up to 3 mm day⁻¹ between 2°S and 2°N throughout the Pacific, relative to the observations, whereas the RZM shows an increase in precipitation by up to 4 mm day⁻¹ in this region, relative to the CTL, indicating that the bias related to the dry tongue is generally eliminated by using the revised ZM convection scheme.

Because the double ITCZ biases are in the Pacific, we will focus on this region in the rest of the paper.

The seasonal cycle of tropical precipitation averaged over 180°–130°W from CMAP, CTL, and RZM are shown in Fig. 2. The observations (Fig. 2a) show that the major ITCZ precipitation is located south of the equator from November to March and north of the equator from March to December. Compared to the CMAP observations, the southern ITCZ in the CTL (Fig. 2b) is present throughout the year, separated by an overly strong dry tongue from the northern ITCZ. In contrast, the seasonal cycle of the ITCZ precipitation produced by the RZM (Fig. 2c) is in better agreement with the observations. In particular, the disappearance of the southern ITCZ in boreal summer and fall agrees well with the CMAP observations. However, the precipitation intensity in both the CTL and the RZM seems excessive in the ITCZ belts.

Corresponding to the difference in ITCZ precipitation, the surface wind field that forces the ocean circulation is also very different between the CTL and the
RZM simulations. Figure 3 shows the wind stress vectors and corresponding magnitudes from the ODAS observations, the CTL, and the RZM, and it shows the differences between the CTL and ODAS and between the RZM and CTL. The ODAS observations (Fig. 3a) show only a single ITCZ located near 7°N that corresponds to the convergence of northeasterly and southeasterly trade winds. In the central South Pacific, westward surface wind dominates because of the counterclockwise deflection of the southeasterly trade wind; therefore, there is no convergence zone. However, associated with the spurious southern ITCZ, the CTL (Fig. 3b) produces an anomalous convergence of northeasterly and southeasterly winds between 5° and 10°S in the central Pacific. Compared to the observations, the simulated easterly wind is also too weak. These deficiencies are more evident in the difference between the CTL and ODAS observations (Fig. 3d). A comparison of Fig. 3c with Figs. 3a,b shows that the southeasterly trade winds simulated in the RZM run (Fig. 3c) is more comparable to the observations. The RZM produces a relative surface wind divergence between 5° and 10°S in the central Pacific compared to the CTL (Fig. 3e) and an increase in easterly wind, resulting in a better simulation of the southeasterly trade winds in the central Pacific.
Accompanying the double ITCZ bias in precipitation and surface wind convergence, the SST field in the CTL also presents significant biases in the tropical Pacific. Figure 4 shows the observed and simulated SST fields and their differences in the Pacific. The observed SST distribution has the warmest SST in the western Pacific warm pool and extends out along the ITCZ and SPCZ. The CTL (Fig. 4b) shows that the warm pool extends eastward zonally between 3° and 12°S to the eastern Pacific such that an anomalous warm-water tongue with maximum SST exceeding 30°C appears in the central and eastern Pacific, which corresponds to the spurious southern ITCZ rainband. Corresponding to the excessive westward extension of the overly dry tongue, the cold tongue in the CTL is also stronger and displaced approximately 10° westward, relative to the observations. The difference in SST between the CTL and the OISST (Fig. 4d) shows that the SST from the CTL in the central Pacific between 180°E and 130°W is warmer by as much as 2°C between 3° and 12°S and colder by as much as 1°C between 2°S and 2°N compared to the observations. There are also large SST biases in the marine stratus region of the west coasts of California and Peru. However, because these large warm biases can be attributed to ocean processes such as coastal upwelling (Large and Danabasoglu 2006), the following analysis will focus only on the SST biases associated with the double ITCZ in the central Pacific. Both the warm SST bias in the southern ITCZ region and the cold SST bias in the cold tongue region are clearly reduced in the RZM run (Fig. 4c). The cold biases in the higher latitudes are also reduced. The improvement can be better seen from the SST difference between the RZM and the CTL runs (Fig. 4e). Overall, the pattern is out of phase with that in Fig. 4d, indicating that the biases in the CTL are alleviated in the RZM run. The SST in the RZM in the central Pacific between 180°E and 130°W is colder by up to 1.5°C between 3° and 12°S and warmer by up to 1°C between 2°S and 2°N compared to the CTL, leading to a nearly 75% decrease in SST bias in the region 3°–12°S and a complete elimination of the cold bias in SST in the region 2°S–2°N.

Not only is the SST simulation improved, but the upper-ocean temperatures and circulations are improved as well. Figure 5 shows the latitude–depth cross section of the upper-ocean temperature and zonal currents in the central Pacific averaged between 180°E and 130°W from the GODAS, CTL, and RZM. Corresponding to the single ITCZ north of the equator, the GODAS observations (Fig. 5a) show a single thermocline ridge near 10°N. Just south of the ridge, there are eastward currents. Farther south, in the region 10°S–5°N, the westward South Equatorial Current (SEC) dominates in the uppermost 30–50 m, whereas the weak South Equatorial Countercurrent (SECC) occurs south of 15°S. In contrast, corresponding to the double ITCZs, the CTL (Fig. 5b) produces a nearly symmetric thermal structure and circulation, with two thermocline ridges located near 9° in each hemisphere. The subsurface water temperature above the 50-m depth in the CTL is warmer by nearly 1°C between 4° and 10°S and colder by

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**FIG. 4.** Annual-mean SST (°C) from (a) OISST, (b) CTL, and (c) RZM and the differences between (d) CTL and OISST and (e) RZM and CTL.
nearly 1°C between 2°S and 2°N compared to the observations, corresponding to the warm SST bias in the southern ITCZ and the cold SST bias in the cold tongue region, respectively. In the latitudes of observed SEC (9°–4°S), the CTL produces a spurious eastward SECC. Conversely, the westward SEC simulated by the CTL between 2°S and 2°N is too strong (by up to 36 cm s⁻¹), which consequently transports more cold water from the eastern Pacific to the central Pacific and could be responsible for the cold bias in that region. On the other hand, the RZM run shows a remarkable improvement in the simulation of the upper-ocean temperature and zonal currents in the central Pacific. Both the spurious thermocline ridge near 9°S and SECC between 9°S and 4°S are eliminated in the RZM (Fig. 5c). The subsurface water temperature above the 50-m depth in the RZM, compared to the CTL, is colder by nearly 1°C between 4° and 10°S and warmer by nearly 1°C between 2°S and 2°N. The westward SEC simulated by the RZM between 2°S and 2°N is also reduced by about 12 cm s⁻¹.

Figure 6 shows the latitude–depth cross section of the upper-ocean salinity in the central Pacific averaged between 180°E and 130°W from Levitus et al. (1998), the CTL, and the RZM. Corresponding to the single ITCZ north of the equator, the Levitus et al. (1998) observation (Fig. 6a) shows a single low-salinity belt near 10°N because of the intense precipitation. In contrast, corresponding to the double ITCZs, the CTL (Fig. 6b) produces two low-salinity belts in the upper ocean located near 10° in each hemisphere, with the southern belt much lower than the northern belt. Because of the decrease of precipitation over the southern ITCZ region, the salinity above the 50-m depth in the RZM increases significantly compared to the CTL, although it still has a low bias compared to observations south of 5°S (Fig. 6c). All these changes make the simulated ocean state in the RZM run in better agreement with observations, indicating that the revised ZM convection scheme can effectively alleviate the biases related to the double ITCZ syndrome in the CCSM3.

The overall changes resulting from the use of the revised ZM scheme, in comparison with the ZM scheme, can be measured by the Taylor diagram (Fig. 7) in the tropics (30°S–30°N). The Taylor diagram is designed such that the radial distance from the origin gives the standard deviation normalized by the observed standard deviation and the azimuth gives the correlation coefficient between the modeled and observed fields. The distance between the model point and the observed point measures the root-mean-square difference of the two fields. Thus, the agreement between the modeled and observed fields is measured by the closeness of the model point to the observation point (denoted as REF) on the x axis. The fields selected in the plot are those commonly used to describe GCM simulations. Different symbols (varying sized upright or inverted triangles and open circles; see figure keys at the upper-left corner) are used to represent range and signs of biases from observations in model simulations. In general, there is an overall improvement from the CTL to the RZM run. The overall weighted root-mean-square-error (RMSE) for the analyzed variables (given at the
upper-right corner) is reduced by about 13% in the RZM compared to the CTL. Except for temperature, the pattern correlation of all the analyzed variables is increased. Significant improvement (>5%) can be found in the pattern correlations of precipitation and short-wave cloud forcing. The bias of Pacific surface wind stress is reduced from 1%–5% to below 1% in the RZM compared to the CTL.

In summary, the above analysis clearly shows that the CCSM3 with the original ZM convection scheme produces typical double ITCZ biases, but these biases are dramatically mitigated when the revised ZM convection scheme is used. It demonstrates that convection parameterization is a deciding factor in simulating a realistic ITCZ–SPCZ system or spurious double ITCZ in the CCSM3. It also demonstrates that the atmospheric model is a dominant component model in the formation of double ITCZ bias in the CCSM3. As a first step to identify the physical mechanisms through which modifications of the convection scheme in the atmospheric model alleviate the double ITCZ and related biases in the upper ocean in the CCSM3, we investigate the impacts of convection schemes on the atmospheric forcing and feedback in the next section.

4. Impacts of convection schemes on atmospheric forcing and feedback

Figure 8 shows the annually averaged precipitation from the uncoupled CAM3 runs CAM3c and CAM3z (counterparts of CTL and RZM, respectively). Compared to the coupled runs, the ITCZ and the SPCZ in both the CAM3 runs are closer to observations. However, an excessive dry tongue penetrating into the warm pool and a southern ITCZ extending into the eastern Pacific are clearly visible in the CAM3c run, whereas, with the revised ZM scheme, no sign of double ITCZ is seen in the CAM3z. In the western Pacific, the CAM3z produces more precipitation than the CAM3c simulation. To understand how the changes in closure assumptions in the Zhang–McFarlane convection scheme may lead to differences in this “initial” precipitation perturbation that is subject to further amplification in the coupled simulation, Fig. 9 shows the dependence of precipitation, 500-mb vertical velocity, and 867-mb vertical velocity (fourth model pressure level above the surface) in the CAM3 simulations on SST in the domain (5°–10°S, 180°E–130°W). The 500-mb vertical velocity measures the convective activity, whereas at 867-mb, taken to be approximately at the planetary boundary layer (PBL) top, vertical velocity measures the PBL convergence. Monthly precipitation and vertical velocity values at each grid point within the domain are binned against SST at 0.5°C intervals. For SSTs below 27.5°C, both the CAM3c and CAM3z show little precipitation, with sinking motion in the midtroposphere and divergence in the PBL. Above 27.5°C, convection in the CAM3c gradually ramps up as SST increases. Correspondingly, the midtroposphere upward motion

Fig. 6. Latitude–depth diagram of annual-mean ocean salinity (g kg⁻¹) averaged over 180°E–130°W for (a) Levitus et al. (1998), (b) CTL, and (c) RZM. Salinity is shaded for values less than 35 g kg⁻¹.
increases and the PBL air becomes convergent. For the CAM3z run, convection does not start to increase significantly until SSTs are greater than 29°C. However, once in the convection regime, much more convection and midtroposphere upward motion are produced for given SST than in the CAM3c. Furthermore, the contribution from the PBL air convergence to the 500-mb upward motion accounts for nearly 80% in the CAM3z, as compared to 50% or less for the CAM3c run, which is seen by comparing the vertical velocities at 500 and 867 mb.

One plausible explanation for this difference in terms of the parameterization closure difference is as follows: In the CAM3c, the ZM convection scheme is closed using CAPE, which is largely determined by local thermodynamic instability associated with the SST. The large-scale circulation only plays a secondary role by supplying moisture through low-level convergence. More moisture supply may produce more condensational heating. However, the heating-induced large-scale circulation has little direct control on further convection. Thus, the precipitation in CAM3c is enhanced gradually as SST increases. For the CAM3z run, the revised ZM scheme is closed using the large-scale generation of CAPE in the free troposphere. Thus, convection is not directly controlled by local instability associated with the SST, resulting in a lower sensitivity of convection to SST. This explains why precipitation in the CAM3z is weaker than that in the CAM3c when SST is colder than 29°C. On the other hand, the new closure induces a stronger interaction between convection and large-scale circulation because it assumes convection is directly controlled by large-scale forcing in the free troposphere. The large-scale circulation drives convection, which in

**Fig. 7.** Taylor diagram evaluating the relative improvements in RZM compared to CTL. Cosine of the angle represents the pattern correlation, and azimuthal distance represents the standard deviation relative to observations. REF point represents zero RMSE compared to observations. For each individual point, upward-pointing triangles represent a positive mean systematic bias and downward-pointing triangles represent a negative mean systematic bias. Solid filled squares are assigned to the experiment with the smallest absolute bias for each metric. (top right) The weighted RMSE relative to CTL is shown. The domain of interest is the tropics (30°S–30°N), unless otherwise indicated.
turn produces additional upward motion that further drives convection. This is similar to the conditional instability of the second kind (CISK)-type of interaction mechanism. In addition, the deep convection in the original ZM scheme is triggered when the CAPE is higher than a threshold ($70 \text{ J kg}^{-1}$), whereas it is only triggered when there is a positive forcing on CAPE from large-scale processes in the free troposphere in the revised ZM scheme. Because CAPE is largely determined by the boundary layer temperature and moisture, this indicates that the trigger condition for deep convection in the revised ZM scheme is more stringent than that in the original scheme. Thus, shallow convection will be more abundant in the CAM3z (Zhang and Mu 2005; Li and Zhang 2008) when the deep convection is suppressed. Because shallow convection is much more efficient in generating low-level convergence (Wu 2003), the enhanced low-level convergence further intensifies convection. These two positive feedback mechanisms lead to stronger convection in the CAM3z once convection is active, when the SST is warmer than $29^\circ\text{C}$.

Referring back to Figs. 4a and 8, SST south of the equator and east of $170^\circ\text{W}$ (southern ITCZ region) is colder than $29^\circ\text{C}$. Because of the lower sensitivity of convection to SST, the CAM3z simulation produces less precipitation than the CAM3c in that region. West of $170^\circ\text{W}$, SST is above $29^\circ\text{C}$. Although convection is active in both models in this region, the CAM3z simulation produces more precipitation than the CAM3c.
because of stronger positive circulation–convection feedback induced by the revised ZM scheme.

Corresponding to the difference in precipitation distribution, surface energy fluxes into the ocean also show significant differences. Figure 10 shows the net surface energy fluxes and their components averaged over 5°–10°S between the two CAM3 runs. Despite having more precipitation east of 170°W, the CAM3c run has more net energy flux into the ocean west of 140°W than the CAM3z run (Fig. 10c). There is about a 30–40 W m⁻² energy flux into the ocean across the Pacific in the CAM3c, whereas the net energy flux in the CAM3z run varies nearly linearly from 10 W m⁻² out of the ocean at the international date line to 50 W m⁻² into the ocean at 115°W. Thus, if the CAM3 is coupled with the ocean model, the net energy flux into the ocean will result in an ocean surface warming in the CAM3c and a relative cooling in the CAM3z west of 140°W. The net shortwave fluxes into the ocean in the CAM3c and CAM3z are about the same east of 140°W, with a maximum near 120°W, and both decrease westward west of 140°W, with a smaller flux in the CAM3z. The reduced shortwave flux in the CAM3z can be attributed to the increase of cloud water path and low-level cloud fraction. The CAM3z run has less shortwave flux gain in the ocean; it also has less longwave loss from the ocean west of 140°W. The latent heat (LH) flux out of the ocean in the CAM3c reaches its maximum near 120°W and decreases westward. On the other hand, in the CAM3z, it reaches its maximum in the warm pool region. In the southern ITCZ region, the latent heat flux out of the ocean in the CAM3z is about 10 W m⁻² more than that in the CAM3c, which can be largely attributed to the stronger easterly wind in the CAM3z. For the net surface energy flux difference, shortwave and latent heat fluxes are the dominant terms. Most of the net energy flux difference between the CAM3c and the CAM3z is due to shortwave radiative flux west of 130°W and latent heat flux east of 130°W. The latent heat flux difference accounts for almost half of the net surface energy flux difference, indicating its potential importance in explaining the difference in ocean–atmosphere coupling.

To further understand the difference between the CAM3c and CAM3z simulations in terms of the atmospheric feedback, we perform a feedback analysis similar to that carried out by Lin (2007). Figure 11 shows the regression coefficients of the net surface energy flux and its components against SST in the CAM3 runs averaged over 5°–10°S. This is a measure of the response of atmospheric forcing to SST change or the potential atmospheric feedback to the ocean. A positive (negative) feedback means that for an increase in SST there is increase (decrease) in net surface energy flux into the ocean, which should further increase (decrease) the SST to amplify (dampen) the initial SST perturbation in a coupled system. The net surface energy flux feedback for the CAM3c is positive west of 140°W and negative
For the CAM3z run, the feedback is negative across the Pacific. Thus, the SST perturbation, which can be produced by the downward net energy flux at surface in the CAM3c, would be subject to further amplification west of 140°W in the coupled model run. On the other hand, the SST perturbations from the surface energy flux would be dampened in the RZM coupled run because of the negative feedback. The breakdown of the net energy flux feedback into shortwave, longwave, sensible heat (SH), and latent heat fluxes shows that shortwave and latent heat flux feedbacks are the dominant components. The longwave flux feedback is in the opposite sign of the shortwave feedback but with much smaller magnitude. The shortwave feedback in the CAM3c is positive west of 160°W, meaning higher SST will lead to more downward shortwave flux into the ocean, and is negative east of 160°W, meaning higher SST will lead to less downward shortwave flux. Compared to the CAM3c, the CAM3z run produces stronger negative feedback west of 120°W, indicating that the SST perturbation will be dampened more effectively than in the CAM3c. For latent heat flux, there is positive feedback west of 130°W for the CAM3c, with a maximum in the warm pool. Thus, higher SST leads to less loss of latent heat from the ocean. For the CAM3z run, the latent heat flux feedback is mostly negative, which means higher SST leads to more loss of latent heat from the ocean.
Why are the shortwave radiative and latent heat flux feedbacks so different between the two runs? Figure 12 plots the regression coefficients of high-, mid-, low-level cloud fractions, as well as the cloud water path against SST for the two simulations. The high-level cloud feedback is positive for both runs, but the peak is shifted from 140°W in the CAM3c to 170°W in the CAM3z. For midlevel clouds, the feedback is also positive for both runs. It increases from close to 0 in the eastern Pacific to about 12% K⁻¹ at 160°W and westward in the CAM3z; it peaks at 140°W in the CAM3c. For low-level cloud fraction, the feedback is more dramatically different. West of 120°W, the feedback is negative for the CAM3c run and positive for the CAM3z. The negative values for cloud fraction feedback in the CAM3c means higher SST will lead to less low-level cloud, thereby permitting more solar radiation into the ocean. The grid-average cloud water path, which is important to cloud reflection of solar radiation, shows dramatic difference between the CAM3c and CAM3z runs as well. Although the feedback for the CAM3c is positive east of the international date line, the values are insignificant (about 10 g m⁻² K⁻¹). On the other hand, for the CAM3z run, the values vary from 10 g m⁻² K⁻¹ at 120°W to nearly 80 g m⁻² K⁻¹ at 160°W and westward. It is well known that trigger condition for deep convection in the original ZM scheme is loose, so deep convection occurs too frequently and lasts too long, resulting in very weak shallow convection in the standard CAM3. Because the original ZM scheme determines convection from CAPE, which largely depends on SST, the shallow convection will be much weaker when the SST is increased due to enhanced deep convection, which consumes more water vapor. This explains why the low-level
cloud fraction feedback is negative west of 120°W in the CAM3c. On the other hand, because the trigger condition for deep convection is more stringent in the revised ZM scheme, shallow convection is more abundant in the CAM3z when deep convection is suppressed. Because the shallow convection scheme in the CAM3 is based on local convective instability, shallow convection will be more active when the SST is increased, resulting in more low-level cloud fraction in the CAM3z. This explains why the revised ZM scheme produces much stronger positive low-level cloud fraction feedback than the original ZM scheme. More active shallow convection, which detains more cloud water into the lower troposphere, contributes to stronger cloud water path feedback in the CAM3z. Thus, both stronger positive low-level cloud and cloud water path feedbacks are responsible to the stronger negative shortwave flux feedback in the CAM3z west of 120°W. In addition, west of 160°W, the CAM3z produces stronger positive cloud feedback than CAM3c, which can be attributed to the stronger convection resulting from enhanced convection–circulation interaction induced by the new closure, leading to stronger negative shortwave flux feedback.

To understand the difference in latent heat flux feedback between the two runs and relate it to convection parameterization, we note that latent heat flux in the model is parameterized by

\[ F_L = -L \rho c_v U (q_s - q), \]  

where \( L \) is the latent heat of evaporation; \( \rho \) is surface air density; \( c_v \) is the turbulent moisture exchange coefficient; \( U \) and \( q \) are the lowest model level wind speed and specific humidity, respectively; and \( q_s \) is the saturation specific humidity at the surface. The negative sign indicates that the heat is out of the ocean, which is consistent with the sign convention of positive downward. Its feedback (i.e., change with SST) can be decomposed into contributions from wind speed and humidity deficit \( \Delta q = (q_s - q) \):

\[ \frac{\partial F_L}{\partial T_s} \approx F_L \left( \frac{\partial U}{U \partial T_s} + \frac{\partial \Delta q}{\Delta q \partial T_s} \right). \]  

Here, we neglected the contribution from the moisture exchange coefficient change with SST, which is small. Figure 13 shows the relative contribution of the wind
speed and humidity deficit feedbacks [the two terms in the parentheses of Eq. (2)]. The humidity deficit increases with SST in both runs, giving positive feedback. However, the strength of the feedback is much smaller than that of the wind speed feedback, indicating that the latent heat flux feedback is largely determined by wind speed. For the CAM3c run, wind speed decreases with SST west of 130°W at relative rates from nearly 0 in the eastern Pacific to about -0.4 K^{-1} at the international date line, which is due to the wind convergence associated with enhanced convection in the central Pacific. The negative wind speed feedback in CAM3z is much weaker than that in the CAM3c, which may be attributed to the lower sensitivity of convection to SST induced by the new closure, as well as the fact that convection is located more toward the western Pacific. Because the latent heat flux out of the ocean increases with wind speed, the stronger negative wind speed feedback results in stronger positive latent heat flux feedback in the CAM3c west of 130°W.

5. Summary and conclusions

This study investigates the impact of convection parameterization on the double ITCZ and associated biases in the NCAR CCSM3 using two versions of the Zhang–McFarlane convection scheme. Comparison of the annual-mean climatologies simulated with the original ZM convection scheme and with the revised ZM scheme clearly shows that all biases related to the double ITCZ syndrome in the CCSM3 with the original ZM scheme are dramatically mitigated when the revised ZM scheme is used. In the cold tongue region (2°S–2°N), the negative bias in precipitation and the cold bias in SST and subsurface water temperature are generally eliminated by using the revised ZM scheme. In the southern ITCZ region (5°–10°S), the positive bias in precipitation and the warm bias in SST in the central Pacific, where the double ITCZ bias is most serious, are reduced by nearly 75% and 66%, respectively. The corresponding spurious thermocline ridge and SECC, as well as the warm bias in subsurface water temperature, are almost entirely eliminated. The southeasterly trade winds and upper-ocean salinity distribution are also improved when the revised ZM scheme is used. The seasonal cycle of precipitation shows that the remaining annual-mean bias in precipitation may be due to the over-prediction of precipitation over the southern ITCZ during boreal winter and spring, when the relatively weak southern ITCZ does occur in observations. These results demonstrate that the convection parameterization plays a critical role in the formation of double ITCZ bias in the CCSM3.

To understand the physical mechanisms through which the modifications of the convection scheme in the atmospheric model alleviate the double ITCZ and related biases in the upper-ocean in the CCSM3, as a first step we investigated the impacts of the convection scheme on the atmospheric forcing and feedback in uncoupled atmospheric model CAM3. It is shown that the CAM3 with the original ZM scheme also produces a double ITCZ bias in precipitation, although it is much weaker than in the coupled model. In contrast, the atmospheric model simulation with the revised ZM scheme shows no sign of double ITCZ in the central Pacific but shows more precipitation in the western Pacific. Diagnostic analyses show that the revised ZM scheme produces less precipitation in the southern ITCZ region because of its lower sensitivity to SST in the SST range observed there, and the positive circulation–convection feedback enhanced by the revised ZM scheme produces more precipitation in the western Pacific. For the atmospheric forcing, the CAM3 with the original ZM scheme produces more net surface energy flux into the ocean west.
of 140°W than that with the revised ZM scheme. This indicates that, when the CAM3 with the original ZM scheme is coupled with the ocean model, the net surface energy flux will result in a relative warming of the ocean surface compared to the revised ZM scheme. For the net energy flux difference, shortwave and latent heat fluxes are the dominant terms. Most of the net energy flux difference is due to shortwave radiative flux west of 130°W and latent heat flux east of 130°W.

The atmospheric feedback analysis shows that the net surface energy flux feedback is negative across the Pacific when the revised ZM scheme is used but is positive west of 140°W and negative east of 140°W when the original ZM scheme is used. A positive feedback means that there is an increase in net surface energy flux into the ocean for an increase in SST, which indicates that the SST perturbations from the surface energy flux will be subject to further amplification in the CTL coupled run but will be dampened in the RZM coupled run. The breakdown of the net energy flux feedback into shortwave, longwave, sensible heat, and latent heat fluxes shows that shortwave and latent heat flux feedbacks are the dominant components.

To summarize, the analysis using the CAM3 simulations identified three factors on the atmospheric side that may contribute to the differences in the coupled simulations: 1) sensitivity of convection to SST, 2) convection–shortwave flux–SST feedback, and 3) convection–wind–evaporation–SST feedback. The original ZM convection scheme is closed using CAPE, which is largely determined by local thermodynamic instability associated with the SST, resulting in a high sensitivity of convection to SST. However, the revised ZM scheme is closed using the large-scale generation of CAPE in the free troposphere. Because convection is not directly controlled by local instability associated with the SST, it leads to a lower sensitivity of convection to SST in the southern ITCZ region. Thus, for a positive SST perturbation in the southern ITCZ region, the original ZM scheme will produce more convection than the revised ZM scheme. For shortwave flux–SST feedback, it is either positive or small negative when the original ZM scheme is used but large negative when the revised ZM scheme is used. Their difference can be largely attributed to the different low-level cloud fraction and cloud water path response to SST and results in more shortwave flux into the ocean, which leads to a relative warming in SST when the original ZM scheme is used compared to the revised ZM scheme. For evaporation feedback, the initial convection perturbation in the central Pacific in the CAM3e leads to weaker surface wind and thus weaker evaporation in and west of convective regions because of the wind convergence. This would provide a positive feedback in a coupled system. On the other hand, when the revised ZM scheme is used, intense convection occurs west of the international date line, the induced easterly convection enhances evaporation, inhibiting the warm SST bias, and thus convection, in the central and eastern Pacific. All three of these factors favor the double ITCZ development in the coupled CTL run more than in the RZM run.

In this paper, the feedback analysis focuses on the atmospheric side by examining the heat input into the ocean and its sensitivity to SST in uncoupled runs driven by prescribed SST. With different heat input and wind stress to the ocean, the ocean circulation will change in response. The roles of the ocean’s heat transport and coupled feedback in the formation of double ITCZ will be investigated in a separate paper (Zhang and Song 2009, submitted to J. Climate).

Acknowledgments. This research was supported by the Biological and Environmental Research Program (BER), the U.S. Department of Energy Grant DE-FG02-03ER63532, and the U.S. National Science Foundation Grant ATM-0601781. The authors thank the reviewers for their valuable comments that helped to improve the manuscript.

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