Improved Simulation of ENSO Periodicity

**Improving ENSO Periodicity Simulation
by Adjusting Cumulus Entrainment in BCC_CSMs**

Bo Lu^{1,2} and Hong-Li Ren^{1,2}

Abstract

The simulation of El Niño-Southern Oscillation (ENSO) phenomenon is a challenging issue for coupled climate models. This study focuses on the ENSO periodicity simulated by Beijing Climate Center Climate System Models (BCC_CSM1.1 and BCC_CSM1.1m) which can reproduce reasonably well ENSO amplitude as observations. However, the major period of ENSO simulated by the BCC_CSMs is around 2.4 years, which is much shorter than that in observations. Compared with other 24 coupled models in Coupled Model Intercomparison Project Phase 5 (CMIP5), BCC_CSMs produce a very unrealistic ENSO peak period. Such a bias in simulating periodicity is suggested as a consequence of the severely underestimated air-sea coupling intensity in BCC_CSMs. To test this hypothesis, a quantitative method is developed to diagnose the linear ENSO frequency. As an effort to improve the ENSO simulation in BCC_CSMs, three experiments are performed with varying entrainment rates in the cumulus convection parameterization scheme of BCC_CSM1.1m. A more realistic ENSO period of about 3.3 years can be generated by the model with an inflated entrainment rate. When the cumulus entrainment is increased by 10%, the ENSO-related convective precipitation will enhance in the equatorial central to eastern Pacific. This anomalous convective heating induces an intensified surface westerly wind stress to the west of the anomalous convection center and as a result, the air-sea coupling intensity becomes larger, which contributes to a longer period of ENSO based on previous theories. In addition, the pronounced eastward extension of ENSO-related surface wind stress could also be the secondary factor to generate a lower frequency of ENSO in BCC_CSMs. Our study proposes a method to reduce the biases in ENSO periodicity simulation and puts more insights into the importance of adjusting atmospheric convection to reproduce ENSO properties in coupled model.

Keywords: ENSO; periodicity; cumulus entrainment; coupled model

1. Introduction

El Niño-Southern Oscillation (ENSO) phenomenon dominates interannual timescale variability in atmosphere-ocean coupled climate system and has broad impacts on global and regional climate anomalies (e.g. Zhang et al., 2011, 2012). In the past three decades, enormous progresses have been made in increasing our understanding of ENSO mechanisms (e.g., Schopf and Suarez, 1988; Battisti and Hirst, 1989; Jin, 1997a, b; Weisberg and Wang, 1997; Picaut et al., 1997; Wang, 2001) and a number of efforts have been put into ENSO prediction (e.g. Latif et al., 1994; Latif et al., 1998; Chen and Cane, 2008; Ren *et al.*, 2014). Nowadays, the coupled global climate models (CGCMs) have been widely employed for predicting ENSO (e.g. Ji et al., 1998; Wang et al., 2002; Jin et al., 2008). Compared with statistical methods (e.g. Graham et al., 1987; Barnston and Ropelewski, 1992; Xue et al., 2000), the potential advantage of using CGCMs is that the ENSO prediction skill can be improved by introducing more realistic physical processes. As the real-time ENSO prediction is still a challenging issue (Barnston et al., 2012) at present, advanced CGCMs with better representations of ENSO properties are needed.

The state-of-the-art CGCMs in Coupled Model Intercomparison Project Phase 5 (CMIP5) show encouraging improvements in simulating ENSO amplitude (Bellenger et al., 2014). The spread of ENSO variability has been constrained in CMIP5 models. The pervasive cold tongue bias (Luo et al., 2005; Rao and Sperber, 2006; Vanniere et al., 2013) is also reduced by 30% from the CMIP3 to CMIP5, and the seasonal locking of ENSO is more realistic (Bellenger et al., 2014). However, ENSO periodicity remains poorly represented. CMIP5 models still have difficulties in capturing the observed spectrum of main ENSO period (Bellenger et al., 2014).

Previous studies have proposed the controlling factors of ENSO periodicity using models with less complexity. One of the key factors is surface wind forcing over the tropical Pacific. Its meridional structure can affect the ENSO period by changing the ocean Rossby radii of deformation (e.g. Kirtman, 1997), and its zonal position can modulate the traveling time of ocean Rossby wave, which thus is also crucial for the ENSO period (e.g. An and Wang, 2000). On the other hand, the importance of air-sea coupling (Hirst, 1986; Jin, 1997b; MacMynowski and Tziperman, 2008), annual cycle (Stuecker et al., 2013), vertical mixing (Syu and Neelin, 2000), wave reflection coefficient (Kang and An, 1998) and non-linearity (Eccles and Tziperman, 2004) are also noted in modulating the ENSO periodicity. However, the above factors are sometimes competing with each other and it is difficult to tune them in state-of-the-art CGCMs in a direct way.

Recent attempts to improve simulations of ENSO amplitude in CGCMs have been made by adjusting the entrainment rate in cumulus convection parameterization. It was found that the simulated ENSO is more sensitive to the changes of atmospheric component than ocean component in CGCMs (Schneider, 2002; Guilyardi et al., 2009). Particularly, the cumulus convection has been suggested to be important in ENSO simulations (Neale et al., 2008). Using a CGCM, Watanabe et al. (2011) showed that a slight reduction of the cumulus convection entrainment will result in a dramatic amplification of the simulated ENSO amplitude. On the contrary, Kim et al. (2011) argued that the magnitude of ENSO tends to be enhanced when the cumulus entrainment is increased in another CGCM. Both of the above studies emphasized the importance of mean state changes with respect to various entrainment parameters in CGCMs. However, almost opposite results were obtained. The divergence of the two results of model experiments presumably indicated that such a convection sensitivity of ENSO amplitude may be

of a dependence upon the CGCMs. Thus, the impact of convective entrainment on ENSO amplitude in CGCMs is still an open question until now. Furthermore, the control of ENSO period by convection and its physical mechanism remain poorly explored.

In this study, we will mainly examine the dependence of ENSO simulation, especially for ENSO periodicity, upon the convective entrainment perturbations by adjusting the cumulus convection parameterization in Beijing Climate Center Climate System Models (BCC_CSMs), and further investigate the improvement of ENSO periodicity simulation based on introducing a quantitative dynamical framework of ENSO. The paper is organized as follows. Observational datasets, CMIP5 experiments, and three numerical experiments using BCC_CSMs are briefly introduced in section 2. ENSO behaviors in CMIP5 simulations are compared in section 3. A quantitative method of estimating ENSO frequency is introduced in section 4. The improvement of ENSO period simulation with varying entrainments is studied in section 5. Summary and conclusion are presented in section 6.

2. Datasets, Models and Experiment Design

a. Observation datasets

NOAA Extended Reconstructed Sea Surface Temperature (SST) V3b (Smith *et al.*, 2008) is used to obtain ENSO features in observations. It provides SST on $2^{\circ} \times 2^{\circ}$ grid from January 1854. Only SST after 1950 is considered in this paper. In addition, Climate Prediction Center Merged Analysis of Precipitation (CMAP) data (Xie and Arkin, 1997) is utilized as well. It covers precipitation from January 1979 on $2.5^{\circ} \times 2.5^{\circ}$ grid. We use the Wave and Anemometer-based Sea-surface Wind (WASWind) dataset to retrieve the surface wind over ocean (Tokinaga and

Xie, 2011). WASWind is constructed from ship observations of wind speed and wind wave height archived in the International Comprehensive Ocean-Atmosphere Data Set (ICOADS). It is available from 1950 to 2011 on $4^{\circ} \times 4^{\circ}$ grid. In this paper, the linear regression method is utilized to obtain the ENSO-induced wind, precipitation and heating changes.

b. CMIP5 models

As suggested by Stevenson et al. (2009), the spectrum of ENSO can be accurately estimated if the length of model simulation exceeds 200 years. Thus, we decided to compare the ENSO period simulations among CMIP5 models from Pre-Industry simulations. They were made available on the Earth System Grid (ESG, <http://pcmdi9.llnl.gov/esgf-web-fe/>). The forcings all remain constant at pre-industrial level (Taylor et al., 2012). The models that provide surface wind, SST, and precipitation are considered in this paper. Most pre-industrial simulations are longer than 300 years (Table 1), which ensures statistical robustness for calculating ENSO period (Wittenberg, 2009). Monthly anomalies are obtained by subtracting the annual cycle for each model. The detailed information is presented in Table 1. It should be noted that two versions (BCC_CSM1.1 and BCC_CSM1.1m) of BCC_CSMs are included in CMIP5. BCC_CSM1.1 is a low-resolution (T42) version, and BCC_CSM1.1m is a median resolution version (T106).

c. BCC_CSMs

BCC_CSM1.1 and BCC_CSM1.1m have been developed for the operational ENSO prediction (Ren et al., 2016) and the participation of CMIP5. Its ocean component is developed by modifying MOM4 from Geophysical Fluid Dynamics Laboratory (Griffies et al., 2004). Its horizontal resolution is $1^{\circ} \times 1^{\circ}$ poleward of 30°N and 30°S , incrementally descending to $1/3^{\circ}$

latitude within 30°N and 30°S. This ocean model has 40 vertical layers and includes dynamic and thermodynamic Sea Ice Simulator (SIS; Winton, 2000). Land surface component is adopted from Atmosphere-Vegetation Interaction Model (AVIM2; Ji et al., 2008). The atmospheric component of BCC_CSMs (BCC_AGCM) is developed from NCAR CAM3 (Collins et al., 2004), with the reference atmosphere and surface pressure modified (Wu et al., 2008). The improvement of physical process from CAM3 is even larger. A new parameterization scheme of cumulus convection (Wu, 2012), a dry adiabatic adjustment scheme and the updated heat exchange process with the ocean surface (Wu et al., 2010) have all been adopted.

d. Design of sensitivity experiments

Cloud profiles and rainfall are very sensitive to the cumulus entrainment scheme. The parameterization for entrainment of the updraft is often a prescribed function (e.g., Zhang and McFarlane, 1995; Tiedtke, 1989). Wu (2012) developed a novel method for estimating the entrainment rate implicitly in BCC_CSMs. Based on total energy conservation, the entrainment rate can be determined by the properties of cloud parcel and environment air:

$$\lambda(z) = \frac{g \frac{T_{v,c}}{\bar{T}_v} + L \frac{dq_c}{dz} + C_p \frac{dT_c}{dz}}{\bar{h} - h_c} \quad (1)$$

where $T_{v,c}$, q_c , T_c and h_c indicate the virtual temperature, water-vapor mixing ratio, temperature and moist static energy of the adiabatic ascending cloud, respectively. $\bar{T}_v$ and $\bar{h}$ are the virtual temperature and moist static energy of the environment air, respectively. Over the West Pacific warm pool, the mean entrainment rate near the cloud base is 0.6 km^{-1} calculated by Eq.(1).

Using this convection parameterization scheme, BCC_AGCM can reproduce the intensity and evolution of several precipitation events (Wu, 2012). Hereafter, control experiment (CTL) represents the experiment using the standard version of BCC_CSM1.1m. Two additional sensitivity experiments are designed to be contrasted with the CTL. During each integration step, the entrainment rate calculated from Eq. (1) is multiplied by a factor, with all other processes unchanged. This factor is set to 0.9 and 1.1 in the sensitivity experiments, respectively, implying that conservation-based entrainment scheme overestimates and underestimates the actual entrainment. It should be noted that these factors are not artificial. In physics, the turbulent entrainment may explain such a potential bias (Houghton and Cramer, 1951). Hereafter, Entr11 and Entr09 represent the experiments with the factors at 1.1 and 0.9, respectively. It could be helpful to perform more experiments by further reducing or increasing the entrainment rate as the previous study of Kim et al. (2011). By tuning the entrainment threshold (Tokiooka parameter) in GFDL-CM2.1, Kim designed four experiments to investigate the ENSO simulations. However, different from GFDL-CM2.1, the entrainment rate in BCC_CSM1.1m is implicitly calculated based on the total energy conservation in the cloud parcel and the environment (Eq. 1; Wu 2012). Thus, the ENSO simulation would be unstable when larger entrainment disturbance is included. This is why we only adjusted entrainment by 10% to represent the turbulent entrainment. All the experiments are run for 44 years after full spin-up.

3. Simulations of ENSO properties in CMIP5 PI-control runs

The ENSO amplitude, as measured by the standard deviation of SST anomalies in Niño3.4 region (5°S-5°N, 170°W-120°W), is well captured in CMIP5 models, as see in Fig. 1. The

amplitude in multi-model ensemble (MME) is 0.87K, very close to the observed value of 0.83K. The spread of simulated ENSO amplitude in CMIP5 is reduced by a factor of 2, compared with CMIP3 simulations (Bellenger et al., 2014). ENSO amplitude as simulated in BCC_CSM1.1 is 0.7K, which is slightly smaller than observation. In contrast, BCC_CSM1.1m reproduces a stronger ENSO than observation. Overall, the bias of ENSO amplitude in BCC_CSM_1.1 (m) is comparable to other CMIP5 models. Fig. 2 shows the patterns of SST standard deviation in observation and CMIP5 simulations. A common feature is that strongest SST variation occurs in central-eastern equatorial Pacific within 5°S-5°N. The westward extension of SST variation is recognized in MME. This is due to the so-called cold tongue bias (Vanniere et al., 2013) which has been much reduced, compared with previous CMIP3 models (Bellenger et al., 2014). For BCC_CSMs, the patterns are reproduced as reasonably well as observations.

Fig. 3 shows the ENSO periods in observations and simulations. ENSO's period varies between 2 and 7 years, with its average around 4 years. Following previous studies (e.g. MacMynowski and Tziperman 2008), we choose 4 years as the observed ENSO period for simplicity. The power spectrum of observed ENSO will be given later in this paper. Unfortunately, unlike amplitude, the improvement of ENSO periodicity simulation is very limited in CMIP5, compared to CMIP3 (Bellenger et al., 2014). The large spread of simulated ENSO period is still very pronounced in CMIP5 MME. Specifically, the period in BCC_CSM1.1 and BCC_CSM1.1m are 2.2 years and 2.4 years, respectively, which are the shortest and second shortest period among all CMIP5 simulations, implying there is an excessively fast ENSO transition in BCC_CSMs. The uncertainties of ENSO periods in BCC_CSMs are estimated by dividing the control runs into 44-year windows, which give us 11 members for BCC_CSM1.1 and 9 members for BCC_CSM1.1m. The period biases are consistent throughout all the samples

(Fig. 3). Thus, ENSO prediction in these models will be highly affected by this unrealistic periodicity. In this paper, we will mainly examine the possible reason of such a period bias in BCC_CSMs and propose a potential method to reduce this bias.

In the equatorial Pacific, the thermocline feedback (anomalous subsurface temperature advected by mean upwelling) and the zonal advective feedback (mean zonal SST gradient advected by anomalous eastward current) are the two major physical processes that lead to ENSO phase transition (An et al., 1999). The non-equilibrium adjustment of zonal mean thermocline depth to equatorial wind stress is the key to form ENSO phase transition (Jin, 1997a, b). An and Wang (2000) coupled the ZC model (Zebiak and Cane 1987) with a statistical atmospheric model, and designed several experiments with various wind stress patterns. They argued that the zonal position of ENSO-related surface wind stress is dominant in simulating ENSO period. A westward shift of wind stress anomalies can reduce the time for ocean Rossby wave to reach the western boundary and thus shorten ENSO period. Fig. 4 evaluates this argument in CMIP5 models. No apparent relationship can be obtained between simulated periods and the zonal positions of ENSO-related surface wind. BCC_CSMs can simulate a moderate zonal position of surface wind, compared with other CMIP5 models. However, the ENSO period is the shortest in BCC_CSMs. It implies that the specific reason for the ENSO period bias may be model dependent.

Previous theoretical studies suggest that air-sea coupling coefficient in tropical Pacific is also a key contributor to ENSO period (Hirst, 1986; Jin, 1997b; MacMynowski and Tziperman, 2008). Fig. 5 compares the air-sea coupling strength among CMIP5 models and observations. Dynamically speaking, the westerly wind anomalies can be generated from zonal SST gradient, which is simply referred to as the air-sea coupling in this manuscript, neglecting the difference of

wind-drag coefficient between models. In observations, the warming of 1K in the eastern Pacific can induce the westerly wind anomaly of 0.6 m/s. This coupling strength is about 0.6m/s per K in terms of MME, although a large spread exists. In contrast, for BCC_CSM1.1(m), the coupling strength is only 0.37 (0.41) m/s per K, which is much weaker than either other CMIP5 models or observation. Such a coupling strength in BCC_CSM1.1(m) ranks the second (third) weakest among 26 CMIP5 models. This implies that the shortest periods in BCC_CSMs are likely attributed to such weak air-sea coupling strengths in these models. More importantly, the bias in ENSO-related surface wind stress can affect the recharge-discharge process that is crucial for ENSO phase transition. In the next section, a quantitative framework will be established to illustrate this argument.

4. Quantitative framework for diagnosing ENSO period

Following recharge-discharge framework (Jin 1997a), the linear tendency equation for SST and thermocline depth anomalies can be written as

$$\begin{cases} \frac{\partial T}{\partial t} = R \times T + F \times h \\ \frac{\partial h}{\partial t} = -\varepsilon \times h - F_t \times T \end{cases} \quad (2)$$

where T and h denote the SST and thermocline depth anomalies, respectively. F and F_t are constant parameters that denote the rate of recharge-discharge process. ε denotes the damping rate of thermocline depth due to the energy leak at western boundary. R is a constant indicating the growth rate of ENSO. Eq. (4) gives the physical processes that affecting the growth rate of ENSO. More details can be found in Jin et al. (2006).

$$R = \mu\beta_u < -\frac{\partial \bar{T}}{\partial x} >_E + \mu\beta_w < -\frac{\partial \bar{T}}{\partial z} >_E + \mu a_h \beta_h < \frac{\bar{w}}{H_m} >_E - \left(\frac{\langle \bar{u} \rangle_E}{L_x} + \frac{-2y \langle \bar{v} \rangle_E}{L_y} + \frac{\langle \bar{w} \rangle_E}{H_m} \right) - \alpha \quad (4)$$

The right hand side of Eq. (4) represents three positive feedbacks and two negative feedbacks which modulate ENSO amplitude. The first three terms are zonal advective feedback, Ekman feedback and thermocline feedback, respectively. The parameters of $\beta_u, \beta_w, \beta_h$ describe the responses of zonal current, vertical motion and thermocline slope to wind stress changes. The parameter of a_h denotes the impact of thermocline changes on subsurface temperature. The terms of $< -\frac{\partial \bar{T}}{\partial x} >_E$ and $< -\frac{\partial \bar{T}}{\partial z} >_E$ indicate the zonal and vertical gradients of climatologic SST. $\bar{w}$ denotes the climatologic upwelling. H_m is the effective depth for vertical advection. Note that every positive feedback term includes the parameter of μ , which represents air-sea coupling strength. μ can be obtained by regressing the wind stress anomaly onto SST changes related to ENSO. The last two terms of Eq. (4) describe the dynamical and thermal damping processes.

Combining Eqs. (2) and (3), we can get an equation with only SST as unknown variable

$$\frac{\partial^2 \langle T \rangle_E}{\partial t^2} + (\varepsilon - \mu \times R_1 + R_2) \frac{\partial \langle T \rangle_E}{\partial t} + (F_t F - \varepsilon \mu \times R_1 + \varepsilon R_2) \langle T \rangle_E = 0 \quad (5)$$

where

$$\text{Positive feedbacks: } R_1 = \beta_u < -\frac{\partial \bar{T}}{\partial x} >_E + \beta_w < -\frac{\partial \bar{T}}{\partial z} >_E + a_h \beta_h < \frac{\bar{w}}{H_m} >_E \quad (6)$$

$$\text{Negative feedbacks: } R_2 = \left(\frac{\langle \bar{u} \rangle_E}{L_x} + \frac{-2y \langle \bar{v} \rangle_E}{L_y} + \frac{\langle \bar{w} \rangle_E}{H_m} \right) + \alpha \quad (7)$$

After assuming a solution as $\langle T \rangle_E = Ae^{i\omega t}$, the growth rate of ENSO can be obtained as $\frac{\mu \times R_1 - R_2 - \varepsilon}{2}$. This growth rate is often regarded as Bjerknes (BJ) instability index (Jin et al., 2006). It has been widely used to diagnose the stability of ENSO amplitude in observations and in coupled simulations (e.g., Kim and Jin, 2010; Lübbecke and McPhaden, 2014).

The linear angular frequency of ENSO can also be obtained from Eq. (5):

$$\varpi_0^2 = F \times F_t - \left(\frac{\mu \times R_1 + \varepsilon - R_2}{2} \right)^2 \quad (8)$$

Eq. (8) gives a novel quantitative method to diagnose ENSO period. This index includes nonlinear terms, which is more complex than BJ index. The term of $\left(\frac{\mu \times R_1 + \varepsilon - R_2}{2} \right)^2$ is often neglected in previous work (Clarke et al., 2007). From Eq. (8), the underestimated air-coupling coefficient, μ , can result in a larger angular frequency and thus a shorter period. It should be noted that a larger μ does not always lead to a longer ENSO period. Nonlinear effect can modulate the $\varpi_0 - T$ relation (Eccles and Tziperman, 2004). In fact, larger μ can result in lower ϖ_0 only when μ is small, and the $\varpi_0 - T$ relation can reverse when μ is large (figure 1 in Jin, 1997b). Compared with the observation and CMIP5 simulations, the air-sea coupling is very weak in BCC_CSMs, explaining the short period bias. In the observation, μ is larger in the mature phase, and this nonlinearity plays a important role for the ENSO asymmetry (An and Jin 2004). We also examine the air-sea coupling in BCC_CSM1.1m for both developing and mature ENSO phases (Fig. 6). It is found that the asymmetry of μ is weak in both phases, and the skewness of Niño3.4 anomaly in BCC_CSM1.1m is only 0.05.

275

276 **5. Entrainment adjustment experiments**

277 In this section, we will design experiments to obtain a more realistic air-sea coupling
 278 strength and ENSO period in BCC_CSM1.1m. Recent studies have suggested that ENSO
 279 properties are sensitive to cumulus entrainment (Kim et al., 2011; Watanabe et al., 2011). The
 280 vertical profiles of mean entrainment rate in BCC_CSM1.1m are shown in Fig. 7. Entrainment
 281 rates are about 0.6 km^{-1} near the cloud base and decrease with height. Compared with the
 282 observation during Kwajalein Experiment (figure 8 in de Rooy et al., 2013), the cumulus
 283 entrainment is underestimated in BCC_CSM1.1m. As an effort to reduce the ENSO period bias
 284 in BCC_CSM1.1m, three experiments are implemented by adjusting the entrainment rate. Fig. 8
 285 shows Morlet wavelet spectra (Morlet et al., 1982) of Niño3.4 SST anomalies in the observation,
 286 CTL, Entr09, and Entr11 experiments. The observation shows a broad spectrum curve centered
 287 around 4 years; however, a sharp peak at 2.7 years is evident in CTL. In other words,
 288 BCC_CSM1.1m produces a fairly regular oscillation of ENSO with a much higher frequency.
 289 Given that the entrainment rate is underestimated in CTL, Entr09 involves a more unrealistic
 290 cumulus process by further reducing the entrainment rate by 10%. As a result, Entr09 still
 291 produces the short ENSO period of 2.5 years. In contrast, when entrainment rate is increased by
 292 10% as simulated in Entr11, a longer period of 3.3 years can be generated. This improvement is
 293 pronounced, although the spectrum in Entr11 is still quite sharp due to the relative short
 294 simulation time. Thus, the increase of cumulus entrainment can reduce the ENSO period bias in
 295 BCC_CSM1.1m. Further, the physical process that works to improve ENSO period simulation in
 296 Entr11 needs to be investigated in depth.

The development of ENSO is accomplished by the vertical advection of anomalous subsurface temperature by the mean upwelling and the zonal advection of mean SST by anomalous current. The former process is so-called "thermocline feedback" (Jin, 1997a, b), and the later process is often referred to as "zonal advective feedback" (Picaut et al., 1997). Zonal advective feedback is crucial in ENSO transition (An et al., 1999). Kim et al. (2011) suggested that zonal gradient of mean SST, a contributor to zonal advective feedback, is sensitive to the cumulus entrainment in GFDL-CM2.1. Thus, we first compare the zonal structure of climatologic SST, thermocline depth and precipitation in CTL, Entr09 and Entr11. As shown in Fig. 9, the zonal mean SST and thermocline depth remains unchanged. The climatologic precipitation in the central Pacific is slightly reduced with larger entrainment in Entr11; however, the changes are limited. We also examine the tropical and global mean SST, precipitation and radiative fluxes with various entrainment rates (Table 2). The results show that the mean state differences are limited among three experiments in BCC_CSM1.1m. The tropical mean SST and precipitation remain unchanged. In fact, the tropical mean precipitation is 3.44 mm/day in BCC_CSM1.1m, which is very close to the observation of 3.37 mm/day. Above all, improved ENSO period simulation is difficult to be attributed to the mean state changes.

Another contributor to zonal advective feedback is the ENSO-induced anomalous eastward current. As suggested by Jin et al. (2006), zonal current anomaly is linearly correlated with the surface wind and the thermocline depth anomalies. The former correlation represents the wind driven current, and the latter correlation indicates the influence from the ocean dynamics since the thermocline depth is crucial for the gravity wave speed. We verify these relationships among the sensitivity experiments in BCC_CSM1.1m. As shown in Fig. 10a, the zonal current anomaly is strongly affected by the surface wind. The correlation coefficient between the zonal current

and the surface wind is 0.7, which is significant at the 99% level under the student- t test. On the other hand, the zonal current anomaly is weakly correlated with the thermocline depth variation (Fig. 10b), implying that the wind-driven effect plays a primary role in BCC_CSM1.1m. Another interesting feature in Fig. 10a is that the response of zonal current to the surface wind remains unchanged in all sensitivity experiments, suggesting that the changes of zonal advective feedback mainly result from the wind response changes in the atmosphere. Thus, we then examine the surface wind response to SST anomalies. This is referred to as the air-sea coupling strength in this paper, neglecting the effect of wind-drag coefficient. Fig. 5 shows that the air-sea coupling strength is significantly weaker in BCC_CSM1.1m than the MME and the observation. After the enhancement of entrainment rate in Entr11, the air-sea coupling strength is intensified to 0.55 m/s per K. Thus, the bias of wind response in BCC_CSM1.1m is reduced. Fig. 11 shows the regressed patterns of surface westerly wind against Niño 3.4 SST index between Entr11 and CTL. The wind response is intensified in Entr11. The maximum westerly wind response increases from 1.1 to 1.3 m/s per K. This result is very similar to the improvement of ENSO period simulation from CCSM3 to CCSM4 (Deser et al., 2012). The ENSO period in CCSM3 is also prolonged due to the strengthening of the wind stress pattern.

This enhancement of air-sea coupling strength in Entr11 can be understood in a classic Gill theory (Gill, 1980). Fig. 12 demonstrates the ENSO-induced precipitation anomalies by regressing the monthly mean precipitation anomalies are regressed against Niño3.4 SST index. In the observation, the precipitation is enhanced over the center Pacific and ITCZ regions during El-Nino events. The pattern of precipitation response in CTL is similar with the observation; however, the strength is underestimated over the central Pacific. When the cumulus entrainment intensity is increased by 10% in the Entr11 experiment, the precipitation anomalies is apparently

enhanced in the central-eastern Pacific along equator. Especially, the precipitation anomalies are increased by 30% over $180^{\circ}\sim 160^{\circ}\text{W}/6^{\circ}\text{S}\sim 2^{\circ}\text{N}$. Fig. 13 shows the difference of the ENSO-related the apparent heat source (Q_1 ; Yanai et al., 1973) between Entr11 and CTL. The heating pattern mainly follows the ENSO-induced convection. The diabatic warming anomaly is evident in the central to eastern equatorial Pacific (Fig. 13), which is consist with the precipitation anomaly pattern (Fig. 12). The diabatic cooling anomaly is observed in the western Pacific due the anomalous subsidence. This heating pattern favors the westerly wind anomalies to the west of the convective center (Gill, 1980), which explains the enhanced air-sea coupling strength in Entr11 (Fig. 11).

As we discussed in section 4, a larger air-sea coupling strength can result in a longer period. Besides, it is also evident that the ENSO-related wind stress exhibits an eastward extension in Entr11. The maximum surface wind locates between 160°E and 160°W in CTL, and the eastern edge of wind response extended to 150°W in Entr11. This eastward extension could result in longer time for Rossby wave to reach the eastern boundary, and a longer period can be generated in terms of delayed oscillation theory (Schopf and Suarez, 1988; Battisti and Hirst, 1989). This could be the secondary factor to produce a lower frequency of ENSO in Entr11.

Although this paper focuses on the ENSO period, the ENSO amplitude changes with various entrainment rates should also be addressed. As shown in Fig. 14, the standard deviation patterns of SST anomalies are very similar between Entr11 and CTL, and no significant changes are observed. Thus, the ENSO amplitude in BCC_CSM1.1m is insensitive to the cumulus entrainment, which is different from previous findings by MIROC5 (Watanabe et al., 2011) and GFDL-CM2.1 (Kim et al., 2011). We further study the physical reason why the ENSO amplitude remains unchanged. Fig. 11 shows that the air-sea coupling strength is intensified in Entr11

experiment, which could lead to stronger ENSO amplitude by strengthening the thermocline feedback, zonal advective feedback and Ekman feedback (Jin et al., 2006). However, on the other hand, larger entrainment rate also results in the intensified convections over the equatorial central to eastern Pacific (Fig. 12). It leads to stronger shortwave radiation damping in Entr11 (Fig. 15), which counteracts the positive dynamical feedbacks. Thus, the ENSO amplitude remains almost unchanged.

6. Summary and discussion

It is found in this study that a relatively short ENSO period of about 2.4 years exists in the BCC_CSM1.1(m) simulations, which is the shortest ENSO period among the CMIP5 models. This ENSO periodicity bias is likely caused by an underestimated air-sea coupling strength in the BCC_CSMs, based on our analysis results. It is shown that the equatorial Pacific surface westerly wind is severely weak during SST warming in the eastern Pacific. This will affect zonal advective feedback and thermocline feedback which are both crucial to the phase transition of ENSO. In this paper, a quantitative metric is introduced to demonstrate the relationship between the air-sea coupling and ENSO periodicity. Three model experiments were designed by varying the cumulus entrainment intensity in BCC_CSM1.1m. We show that with the cumulus entrainment increasing by 10%, the surface wind anomaly can be enhanced, and then the simulated ENSO period can be evidently extended to 3.3 years, closer to the observation.

The physical processes with larger cumulus entrainment are also studied. ENSO-related convective precipitation is enhanced in 180°~160°W along the equator. The convective heating favors an intensified westerly wind at the lower troposphere. Thus, a stronger atmospheric

response can be generated with the same magnitude of SST variability associated with ENSO. Then, a more realistic ENSO period can be simulated based on the quantitative metric as developed in section 4. Indeed, the control of cumulus entrainment on ENSO properties may be model dependent. Previous studies by using GFDL-CM2.1 (Kim et al., 2011) and MIROC5 (Watanabe et al., 2011) suggested that cumulus entrainment is quite important for ENSO amplitude, but not for period, by changing the Pacific mean state. Here, we show that the ENSO periodicity in BCC_CSM1.1m is very sensitive to cumulus convective entrainment. Interestingly, unlike GFDL-CM2.1 and MIROC5, the tropical mean (Table 2) and zonal mean structure (Fig. 9) of the radiation balance, climatologic SST, thermocline depth and precipitation remains almost unchanged with various entrainment rates. Instead, the stronger ENSO-induced atmospheric response plays an important role in reducing the ENSO periodicity bias of CGCMs.

As an effort to improve the robustness of our results, we extend Entr11 by 300 years and split the total run into 6 ensemble members. It is found that the prolonged ENSO period is consistent among each member (Fig. 3). This work argues for the importance of atmospheric convection on ENSO property simulations. An urgent need to produce a more realistic cumulus entrainment is especially essential. More efforts on cloud resolution models and observations (Luo et al., 2010) would be a potential way.

In addition, for the ENSO period bias issue, we need to consider another possibility that the about 2.4-year period in BCC_CSM1.1(m) may be related to the so-called quasi-biennial ENSO mode emerging in the air-sea coupled equatorial Pacific system, corresponding to the non-conventional type of ENSO (Ren and Jin, 2013; Ren et al., 2013). Along this line, it will be interestingly worthy studying that the adjustment of atmospheric cumulus entrainment process may change the ENSO regime in terms of stability changing of intrinsic ENSO modes.

411

412 **Acknowledgements.** This work is jointly supported by China Meteorological Special Program
413 (GYHY201506013), National Key Research Program and Development of China
414 (2016YFA0602100), and National Science Foundation (41605116).

415

416 **Conflict of Interest** The authors declare no conflict of interest.

417

418 **References:**

419 An, S.-I., Jin, F.-F., 2004. Nonlinearity and asymmetry of ENSO. *J. Clim.* 17, 2399–2412.

420 An, S.-I., Jin, F.-F., Kang, I.-S., 1999. The role of zonal advection feedback in phase transition
421 and growth of ENSO in the Cane-Zebiak model. *J. Meteor. Soc. Japan* 77, 1151–1160.

422 An, S.-I., Wang, B., 2000. Interdecadal change of the structure of the ENSO mode and its impact
423 on the ENSO frequency. *J. Clim.* 13, 2044–2055.

424 Barnston, A.G., Ropelewski, C.F., 1992. Prediction of ENSO episodes using canonical
425 correlation analysis. *J. Clim.* 5, 1316–1345.

426 Barnston, A.G., Tippett, M.K., L'Heureux, M.L., Li, S., DeWitt, D.G., 2012. Skill of Real-Time
427 Seasonal ENSO Model Predictions during 2002–11: Is Our Capability Increasing? *Bull. Amer.*
428 *Meteor. Soc.* 93, 631–651.

- 429 Battisti, D., Hirst, A., 1989. Interannual variability in a tropical atmosphere-ocean model:
430 Influence of the basic state, ocean geometry and nonlinearity. *J. Atmos. Sci.* 46, 1687–1712.
- 431 Bellenger, H., Guilyardi, E., Leloup, J., Lengaigne, M., Vialard, J., 2014. ENSO representation in
432 climate models: from CMIP3 to CMIP5. *Clim. Dyn.* 42, 1999–2018.
- 433 Chen, D., Cane, M.A., 2008. El Niño prediction and predictability. *J. Comput. Phys.* 227, 3625–
434 3640.
- 435 Clarke, A.J., Van Gorder, S., Colantuono, G., 2007. Wind stress curl and ENSO
436 discharge/recharge in the equatorial Pacific. *J. Phys. Oceanogr.* 37, 1077–1091.
- 437 Collins, W.D., Rasch, P.J., Boville, B.A. et al. 2004. Description of the NCAR Community
438 Atmosphere Model (CAM3.0). NCAR/TN-464+STR, 214 pp.
- 439 de Rooy, W.C., Bechtold, P., Frohlich, K., Hohenegger, C., Jonker, H., Mironov, D., Siebesma,
440 A.P., Teixeira, J., Yano, J.-I., 2013. Entrainment and detrainment in cumulus convection: an
441 overview. *Q. J. R. Meteorol. Soc.* 139, 1–19.
- 442 Deser, C., et al., 2012. ENSO and Pacific decadal variability in the Community Climate System
443 Model version 4. *J. Clim.* 25, 2622–2651
- 444 Eccles, F., Tziperman, E., 2004. Nonlinear effects on ENSO’s period. *J. Atmos. Sci.* 61, 474–482.
- 445 Gill, A.E., 1980. Some simple solutions for heat induced tropical circulations. *Quart. J. R.*
446 *Meteor. Soc.* 106, 447–462.
- 447 Graham, N.E., Michaelsen, J., Barnett, T.P., 1987. An investigation of the El Niño-Southern
448 oscillation cycle with statistical models. 2. Model results. *J. Geophys. Res.* 92, 14271–14289.

- 449 Griffies, S.M., Harrison, M.J., Pacanowski, R.C., Rosati, A., 2004. A Technical Guide to MOM4,
450 NOAA/Geophysical Fluid Dynamics Laboratory, Princeton, USA, 337 pp.
- 451 Guilyardi, E., Wittenberg, A., Fedorov, A. et al., 2009. Understanding El Niño in Ocean–
452 Atmosphere General Circulation Models: Progress and Challenges. *Bull. Amer. Meteor. Soc.*
453 90, 325–340.
- 454 Hirst, A.C., 1986. Unstable and damped equatorial modes in simple coupled ocean–atmosphere
455 models. *J. Atmos. Sci.* 43, 606–630.
- 456 Houghton, H., Cramer, H., 1951. A theory of entrainment in convective currents. *J. Meteorol.* 8,
457 95–102.
- 458 Ji, J., Huang, M., Li, K., 2008. Prediction of carbon exchanges between China terrestrial
459 ecosystem and atmosphere in 21st century. *Sci. China (Ser. D)*, 51(6), 885-898.
- 460 Ji, M., Behringer, D.W., Leetmaa, A., 1998. An improved coupled model for ENSO prediction
461 and implications for ocean initialization. Part II: The coupled model. *Mon. Weather Rev.* 126,
462 1022–1034.
- 463 Jin, E.K., Kinter, J.L. III, Wang, B. et al., 2008. Current status of ENSO prediction skill in
464 coupled ocean–atmosphere models. *Clim. Dyn.* 31, 647–664.
- 465 Jin, F.-F., 1997a. An equatorial ocean recharge paradigm for ENSO. Part I: Conceptual model. *J.*
466 *Atmos. Sci.* 54, 811–829.
- 467 Jin, F.-F., 1997b. An Equatorial Ocean Recharge Paradigm for ENSO. Part II: A Stripped-Down
468 Coupled Model. *J. Atmos. Sci.* 54, 830–847.

- 469 Jin, F.-F., Kim, S.-T., Bejarano, L., 2006. A coupled-stability index for ENSO. *Geophys. Res.*
 470 *Lett.* 33, L23708, doi:10.1029/2006GL027221.
- 471 Kang, I.S., An, S.-I., 1998. Kelvin and Rossby wave contributions to the SST oscillation of
 472 ENSO. *J. Clim.* 11, 2461–2469.
- 473 Kim, D., Jang, Y.-S., Kim, D.-H. et al., 2011. El Niño–Southern Oscillation sensitivity to
 474 cumulus entrainment in a coupled general circulation model. *J. Geophys. Res.* 116, D22112,
 475 doi:10.1029/2011JD016526.
- 476 Kim, S.-T., Cai, W., Jin, F.-F., Yu, J.-Y., 2014. ENSO stability in coupled climate models and its
 477 association with mean state. *Clim. Dyn.*, 42, 3313–3321.
- 478 Kim, S.-T., Jin, F.-F., 2010. An ENSO stability analysis. Part II: results from the twentieth and
 479 twenty-first century simulations of the CMIP3 models. *Clim. Dyn.* 36, 1609–1627.
- 480 Kirtman, B.P., 1997. Oceanic Rossby wave dynamics and the ENSO period in a coupled model.
 481 *J. Clim.* 10, 1690–1704.
- 482 Latif, M., Anderson, D., Barnett, T., Cane, M.A., Kleeman, R., Leetmaa, A., O’Brien, J., Rosati,
 483 A., Schneider, E., 1998. A review of the predictability and prediction of ENSO. *J. Geophys.*
 484 *Res.* 103, 14375–14394.
- 485 Latif, M., Barnett, T.P., Cane, M.A., Flügel, M., Graham, N.E., von Storch, H., Xu, J.-S., Zebiak,
 486 S.E., 1994. A review of ENSO prediction studies. *Clim. Dyn.* 9, 167–179.
- 487 Lübbecke, J.F., McPhaden, M.J., 2014. Assessing the Twenty-First-Century Shift in ENSO
 488 Variability in Terms of the Bjerknes Stability Index. *J. Clim.* 27, 2577–2587.

- 489 Luo, J.-J., Masson, S., Roeckner, E., MAdec, G., Yamagata, T., 2005. Reducing climatology bias
490 in an Ocean-Atmosphere CGCM with improved coupling physics. *J. Clim.* 18, 2344-2360.
- 491 Luo, Z.J., Liu, G.Y., Stephens, G.L., 2010. Use of A Train data to estimate convective buoyancy
492 and entrainment rate. *Geophys. Res. Lett.* 37, L09804, doi:10.1029/2010GL042904.
- 493 MacMynowski, D.G., Tziperman, E., 2008. Factors Affecting ENSO's Period. *J. Atmos. Sci.* 65,
494 1570–1586.
- 495 Morlet, J., Arens, G., Fourgeau, E., Glard, D., 1982. Wave propagation and sampling theory—
496 Part I: Complex signal and scattering in multilayered media. *Geophys.* 47, 203-221.
- 497 Neale, R.B., Richter, J.H., Jochum, M., 2008. The impact of convection on ENSO: From a
498 delayed oscillator to a series of events. *J. Clim.* 21, 5904–5924.
- 499 Picaut, J., Masia, F., du Penhoat, Y., 1997. An advective-reflective conceptual model for the
500 oscillatory nature of the ENSO. *Science* 277, 663-666.
- 501 Rao, A.K., Sperber, K.R., 2006. ENSO simulation in coupled ocean-atmosphere models: Are the
502 current models better? *Clim. Dyn.* 27, 1-15.
- 503 Ren, H.-L., Jin, F.-F., 2013. Recharge oscillator mechanisms in two types of ENSO. *J. Clim.* 26,
504 6506-6523.
- 505 Ren, H.-L., Jin, F.-F., Stuecker, M.F., Xie, R., 2013. ENSO regime change since the late 1970s
506 as manifested by two types of ENSO. *J. Meteor. Soc. Jap.* 91(6), 835-842.
- 507 Ren, H.-L., Liu, Y., Jin, F.-F., Yan, Y.-P., Liu, X.-W., 2014. Application of the analogue-based
508 correction of errors method in ENSO prediction. *Atmos. Oceanic Sci. Lett.* 7(2), 157–161.

- 509 Ren, H.-L., Liu, Y., Zuo, J., Lu, B., Tian, B., Jin, F.-F., Wan, J., 2016.
 510 The new generation of ENSO prediction system in Beijing Climate Center and its forecast for
 511 2014/16 super El Niño event. *Meteorological Monthly*, 42(5), 521-531.
 512 doi:10.7519/j.issn.1000-0526.2016.05.001. (in Chinese).
- 513 Schneider, E.K., 2002. Understanding differences between the equatorial Pacific as simulated by
 514 two coupled GCMs. *J. Clim.* 15, 449–469.
- 515 Schopf, P.S., Suarez, M.J., 1988. Vacillations in a coupled ocean–atmosphere model. *J. Atmos.*
 516 *Sci.* 45, 549–566.
- 517 Smith, T.M., Reynolds, R.W., Peterson, T.C., Lawrimore, J., 2008. Improvements NOAAs
 518 Historical Merged Land–Ocean Temp Analysis (1880–2006). *J. Clim.* 21, 2283–2296.
- 519 Stevenson, S., Fox-Kemper, B., Jochum, M., Rajagopalan, B., Yeager, S.G., 2010. ENSO model
 520 validation using wavelet probability analysis. *J. Clim.* 23, 5540–5547.
- 521 Stuecker, M.F., Timmerman, A., Jin, F.-F., McGregor, S., Ren, H.-L., 2013. A combination
 522 mode of the annual cycle and the El Niño/Southern Oscillation. *Nat. Geosci.* 6, 540-544.
- 523 Syu, H.H., Neelin, J.D., 2000. ENSO in a hybrid coupled model. Part I: Sensitivity to physical
 524 parameterizations. *Clim. Dyn.* 16, 19–34.
- 525 Taylor, K.E., Stouffer, R.J., Meehl, G.A., 2012. An overview of CMIP5 and the experiment
 526 design. *Bull. Amer. Meteor. Soc.* 93 (4), 485–498, doi:10.1175/BAMS-D-11-00094.
- 527 Tiedtke, M., 1989. A comprehensive mass flux scheme for cumulus parameterization in large-
 528 scale models. *Mon. Wea. Rev.* 117, 1779–1800.

- 529 Tokinaga, H., Xie, S.P., 2011. Wave and Anemometer-based Sea-surface Wind (WASWind) for
530 climate change analysis. *J. Clim.* 24, 267-285.
- 531 Vanniere, B., Guilyardi, E., Madec, G., Doblas-Reyes, F.J., Woolnough, S., 2013. Using
532 seasonal hindcasts to understand the origin of the equatorial cold tongue bias in CGCMs and
533 its impact on ENSO. *Clim. Dyn.* 40, 963-981.
- 534 Wang, C., 2001. A unified oscillator model for the El Niño-Southern Oscillation. *J. Clim.* 14, 98-
535 115.
- 536 Wang, G., Kleeman, R., Smith, N., Tseitkin, F., 2002. The BMRC coupled general circulation
537 model ENSO forecast system. *Mon. Weather Rev.* 130, 975–991.
- 538 Watanabe, M., Chikira, M., Imada, Y., Kimoto, M., 2011. Convective control of ENSO
539 simulated in MIROC, *J. Clim.* 24, 543-562.
- 540 Weisberg, R.H., Wang, C., 1997. A western Pacific oscillator paradigm for the El NiñoSouthern
541 Oscillation. *Geophys. Res. Lett.* 24, 779-782.
- 542 Winton, M., 2000. A reformulated three-layer sea ice model. *J. Atmos. Ocean Technol.* 17, 525-
543 531.
- 544 Wittenberg, A.T., 2009. Are historical records sufficient to constrain ENSO simulations?
545 *Geophys. Res. Lett.* 36, L12702. doi:10.1029/2009GL038710.
- 546 Wu, T., 2012. A mass-flux cumulus parameterization scheme for large-scale models: Description
547 and test with observations. *Clim. Dyn.* 38, 725-744.

- 548 Wu, T., Yu, R., Zhang, F., 2008. A modified dynamic framework for the atmospheric spectral
549 model and its application. *J. Atmos. Sci.* 65, 2235-2253.
- 550 Wu, T., Yu, R., Zhang, F., 2010. The Beijing Climate Center atmospheric general circulation
551 model: Description and its performance for the present-day climate. *Clim. Dyn.*, 34, 123-147.
- 552 Xie, P., Arkin, P.A., 1997. Global precipitation: A 17-year monthly analysis based on gauge
553 observations, satellite estimates, and numerical model outputs. *Bull. Amer. Meteor. Soc.* 78,
554 2539–2558.
- 555 Xue, Y., Leetmaa, A., Ji, M., 2000. ENSO prediction with Markov models: The impact of sea-
556 level. *J. Clim.* 13, 849-871.
- 557 Yanai, M., Esbensen, S., Chu, J., 1973. Determination of bulk properties of tropical cloud
558 clusters from large-scale heat and moisture budgets. *J. Atmos. Sci.*, 30, 612–627.
- 559 Zhang, G.J., McFarlane, N.A., 1995. Sensitivity of climate simulations to the parameterization of
560 cumulus convection in the Canadian Climate Center general circulation model. *Atmos. Ocean.*
561 33, 407–446.
- 562 Zhang, W., Jin, F.-F., Li, J., Ren, H.-L., 2011. Contrasting impacts of two-type
563 El Niño over the western north Pacific during boreal autumn. *J. Meteor. Soc. Jap.*, 89(5), 563-
564 569.
- 565 Zhang, W., Jin, F.-F., Ren, H.-L., Li, J., Zhao, J.X., 2012.
566 Differences in teleconnection over the North Pacific and rainfall shift over the
567 USA associated with two types of El Niño during boreal autumn. *J. Meteor. Soc. Japan.*, 90(4),
568 535-552.

569 Zebiak, S.E., Cane, M.A., 1987. A Model El Niño–Southern Oscillation. Mon. Wea. Rev. 115,
570 2262–2278.

571

Table 1 The length of PI-control runs with CMIP5 models used in this paper

Models	Length of Runs (Yr)	Models	Length of Runs (Yr)
BCC_CSM1.1	500	GISS-E2-R-CC	251
BCC_CSM1.1m	400	HadGEM2-CC	241
BNU-ESM	400	HadGEM2-ES	575
CanESM2	996	inmcm4	500
CMCC-CESM	277	IPSL-CM5A-MR	300
CMCC-CM	330	IPSL-CM5B-LR	300
CMCC-CMS	500	MIROC4h	100
CNRM-CM5	850	MIROC5	700
FGOALS-s2	501	MIROC-ESM	531
GFDL-ESM2G	500	MPI-ESM-LR	1000
GISS-E2-H	1770	MPI-ESM-MR	1000
GISS-E2-H-CC	251	MPI-ESM-P	1156
GISS-E2-R	1200	MRI-CGCM3	500

Table 2 The global and tropical mean (30°S~30°N) of climatologic net longwave flux at surface (FLNS), net solar flux at surface (FSNS), outgoing longwave flux at the top of atmosphere (OLR), surface upward latent heat flux (LHFLX), surface upward sensible heat flux (SHFLX), convective precipitation (PRECC), stable precipitation (PRECL) and sea surface temperature (SST) in Entr09, CTL and Entr11 experiments.

	Entr09	CTL	Entr11
FLNS_global (W/m ²)	56.41	56.50	56.51
FLNS_tropics (W/m ²)	61.38	61.54	61.58
FSNS_global (W/m ²)	132.93	132.93	133.11
FSNS_tropics (W/m ²)	206.60	206.87	207.16
OLR_global (W/m ²)	225.18	225.01	225.27
OLR_tropics (W/m ²)	255.53	255.74	256.07
LHFLX_global (W/m ²)	65.56	65.38	65.60
LHFLX_tropics (W/m ²)	114.65	114.78	115.03
SHFLX_global (W/m ²)	11.04	11.04	11.04
SHFLX_tropics (W/m ²)	22.44	22.44	22.40
PRECC_global (mm/day)	0.92	0.91	0.92
PRECC_tropics (mm/day)	2.00	2.00	2.00
PRECL_global (mm/day)	1.55	1.55	1.55
PRECL_tropics (mm/day)	1.43	1.44	1.44
SST_global (K)	286.63	286.47	286.54
SST_tropics (K)	300.80	300.79	300.83

Figures

Fig. 1. The standard deviation ($^{\circ}\text{C}$) of Niño 3.4 index in ERSST, MME, BCC_CSM1.1 and BCC_CSM1.1m. The error bar for MME indicates one standard deviation of model spread.

Fig. 2. The patterns of standard deviations of SST anomalies ($^{\circ}\text{C}$) in (a) ERSST, (b) CMIP5 MME, (c) BCC_CSM1.1 and (d) BCC_CSM1.1m.

Improved Simulation of ENSO Periodicity

Fig. 3. The peak periods (Year) of Niño 3.4 index in ERSST, MME, BCC_CSM1.1, BCC_CSM1.1m and Entr11. The error bar for MME indicates one standard deviation of multi-model spread. The error bars for BCC_CSMs indicate the period spread of subsets divided from the total run with 44-year window.

Fig. 4. Scatter plot between ENSO period (y-axis, unit: year) and the longitudinal position (x-axis) of ENSO-related surface westerly wind. BCC_CSM1.1, BCC_CSM1.1m and Entr11 are indicated by green, red and blue circles, respectively.

Improved Simulation of ENSO Periodicity

Fig. 5. The air-sea coupling strength (m/s per K) in ERSST, MME, BCC_CSM1.1, BCC_CSM1.1m and Entr11. The error bar for MME indicates one standard deviation of model spread.

Fig. 6. The scatter plot of the surface wind (120°E~160°W, 5°S~5°N) against the Niño3.4 anomaly in the PI-control run by BCC-CSM1.1m.

Fig. 7. The vertical profiles of entrainment rates (m⁻¹) over West Pacific warm pool (110°E~160°E; 10°S~10°N) in CTL (black curve), Entr11 (red curve) and Entr09 (green curve).

Fig. 8. The wavelet power spectra in CTL (black curve), Entr09 (blue curve), Entr11 (red curve), and ERSST (yellow curve).

Fig. 9. Zonal structures for (a) the annual mean climatological SST (K), (b) precipitation (mm/day) and (c) mean thermocline depth (m) in CTL (black curve), Entr09 (blue curve) and Entr11 (red curve). All variables are averaged between 10°S and 10°N.

Fig. 10. The scatter plots of the zonal current anomalies against (a) the surface wind speed and (b) thermocline depth in CTL (black dots), Entr09 (blue dots) and Entr11 (red dots) experiments. The linear regression lines are also given.

Improved Simulation of ENSO Periodicity

Fig. 11. Monthly mean surface zonal wind (m/s per K) regressed against Niño 3.4 indices in (a) CTL and (b) Entr11, as well as (c) the difference of regressed wind patterns between Entr11 and CTL.

Improved Simulation of ENSO Periodicity

Fig. 12. Monthly mean precipitation (mm/day per K) regressed against Niño 3.4 indices in (a) the observation and (b) CTL, as well as (c) the difference of regressed wind patterns between Entr11 and CTL.

Fig. 13. The difference of the ENSO-induced apparent heat source (W/m^2 per K, Yanai 1973) between Entr11 and CTL.

Improved Simulation of ENSO Periodicity

Fig. 14. The patterns of standard deviations of SST anomalies ($^{\circ}\text{C}$) in (a) CTL, (b) Entr11 and (c) Entr11 minus CTL.

Fig. 15. The difference of the ENSO-induced net shortwave radiation at surface (W/m^2 per K) between the Entr11 and CTL.