An Improved Atmospheric Component of Zebiak-Cane Model for Simulating ENSO Winds

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Abstract

The atmospheric component of the Zebiak–Cane (ZC) coupled model is a simple elegant framework that plays an instrumental role in the investigation of the fundamental coupled dynamics of the tropical ocean–atmosphere interaction and the El Niño/Southern Oscillation (ENSO) phenomenon. We attempt two simple thermal and dynamical modifications to reduce apparently spurious wind activities simulated by this atmospheric model in the eastern tropical Pacific under ENSO sea surface temperature (SST) anomaly forcing. First, motivated by established observational evidences, we manipulate the anomalous convective heating to be dependent on the sum of mean SST and SST anomalies and the low-level convergence. Second, we add an ad hoc parameterization for the effect of convective momentum transport (CMT) into the zonal momentum balance.

By adding a background-dependent efficiency factor into convective heating and further adjusting the convective feedback parameter used in the original model, we are able to substantially reduce the wind bias in the tropical Pacific, especially in the eastern tropical Pacific. Meanwhile, with a simple CMT parameterization, it compensates an over reduction in the central Pacific wind due to the first modification. By comparing the simulated wind anomalies in our modified version with that from the original version and with the observation, we find that the equatorial wind response to observation SST anomaly forcing for the period 1982–2010 has been improved in terms of amplitudes and spatial patterns. We also show that these improvements are carried over to ZC coupled model as well when we add these modifications to coupled simulations.

Keywords Zebiak-Cane model; ENSO wind anomalies; convective heating; convective momentum transport
1. Introduction

Capturing the key processes that determine the tropical atmospheric circulation responses to sea surface temperature (SST) anomalies is essential in studying the fundamental dynamics of the tropical ocean–atmosphere interaction. Utilizing the simple linear dynamical framework of the forced tropical atmospheric waves (Matsuno 1966; Webster 1972; Gill 1980), or in short, the MG-type model with the specified atmospheric forcing, Zebiak (1982, 1984) succeeded in simulating equatorial surface wind anomalies of El Niño/Southern Oscillation (ENSO) by replacing the specified steady atmospheric convective heating by a simple parameterization of convective heating related to the SST and convergence anomalies. Moreover, this simple atmospheric model was successfully coupled with an ocean model of the same order of complexity, forming the now famous Zebiak–Cane (ZC) coupled model (Zebiak and Cane 1987, ZC87 hereafter), which was the first dynamical model that successfully forecasted an ENSO event (Cane et al. 1986) and had become a cornerstone in ENSO theoretical studies.

Around the same time, Lindzen and Nigam (1987, LN87 hereafter) proposed a very different approach to simulate surface winds. The LN87 model had no explicit parameterizations of convective heating anomalies. It focused on the dynamical balance in the boundary layer, and the effect of SST anomaly forcing is introduced by relating the surface pressure gradient directly to the SST gradient. Nonetheless, it had to take further consideration of the role of convection through the entrainment in the mass balance. Therefore, Neelin (1989) and Battisti et al. (1999) later demonstrated that the LN87 model is in fact equivalent to the MG-type model with some rearrangements and rescaling of the model equations. Nevertheless, Wang and Li (1993), for instance, took into consideration both the effect of the so-called surface SST gradient forcing and the convective heating by including an explicitly boundary layer model together with explicit parameterizations of convective heating in a slightly more complex formulation of the atmospheric model.

Even though quite different dynamical frameworks were adopted, the simulations from these relatively simple models were roughly in agreement with observation. For instance, the ZC model simulates fairly realistic wind anomalies in the tropical central Pacific during an ENSO event. However, it still exhibits major bias, as noted by Zebiak (1986). As an illustration, we used the observed SST anomalies (November–February mean) of three categories of mature ENSO events to force the original ZC atmospheric model and compare the simulated and observed atmospheric responses.

Here, the three ENSO categories are defined as the cold tongue (CT) El Niño, warm pool (WP) El Niño, and La Niña events. The CT El Niño contains the 1982/83, 1991/92, and 1997/98 events, and the WP El Niño includes the 1994/95, 2002/03, 2004/05, 2006/07, and 2009/10 events. This classification of two types of El Niño events follows that of Kug et al. (2009) whose definition were based on the values of Niño 3 and Niño 4 indices. The La Niña events are defined by Niño 3.4 index following the definition of Trenberth (1997), and it includes the events of 1984/85, 1985/86, 1988/89, 1995/96, 1999/2000, 2000/01, 2007/08, and 2008/09. Note that the multi-year La Niña events lasting for two winters are considered as two consecutive events.

As shown in Fig. 1, the simulated ENSO wind anomalies from the original atmospheric model have some large wind bias, especially in the eastern Pacific (Figs. 1c, g, k). In contrast with the observations, the model overestimates the easterly (westerly) anomalies in the eastern equatorial Pacific during El Niño (La Niña) conditions (Figs. 1d, h, l). Zebiak (1990) attributed this eastern Pacific “problem” to the parameterization of the convective heating anomalies, which was simply proportional to SST anomalies pattern without fully considering the fact that convection hardly occurs in the CT region except during strong El Niño. As a result, the simulated convective heating patterns bear great similarity with the sea surface temperature anomaly (SSTA) pattern (Figs. 2d–f), whereas the observed heating patterns appear to be limited by the background CT temperature (Figs. 1, 2) and has a narrower meridional extent than those in the simulation. These biases in anomalous heating pattern tend to generate strong projections onto the Kelvin waves and thus overactive easterly (westerly) wind anomalies in the El Niño (La Niña) condition. Therefore, the biases in the convective heating pattern are partly responsible for the biases in the wind simulation.

The observational analyses have shown that the convective momentum transport (CMT) plays an important role in the atmospheric momentum budget (Stevens 1979; Wu and Yanai 1994, Tung and Yanai 2002; Dima et al. 2005). Modeling results from a general circulation model have shown that the inclusion of CMT improved the simulating skill of ENSO.
wind anomalies (Kim et al. 2008). Since the ZC atmospheric model omitted this process, in this study, we attempt to add a simple parameterization of CMT into the ZC model to examine its effect on ENSO wind simulation.

This paper is organized as follows. The model modifications are described in Section 2. The method adopted to adjust the heating parameters is discussed in Section 3. The improvements of our modified model in simulating ENSO winds are summarized in Sections 4 and 5. The physical interpretations of the thermal and dynamical effects of added processes are given in Section 6. Summary and discussion are given in Section 7.

2. Model description

The atmospheric component of the ZC model is the reduced-gravity shallow water model on an equatorial beta plane. The original governing equations of the atmospheric component of the ZC model are as follows:

\[ \varepsilon \nabla^2 u^n - \beta_0 \nu \Delta u^n = -(p^n/\rho_0)_x, \]

\[ \varepsilon \nabla^2 v^n + \beta_0 \nu \Delta v^n = -(p^n/\rho_0)_y, \]

\[ \varepsilon (p^n/\rho_0) + c_a^2 [(u_a^n)_x + (v_a^n)_y] = -\dot{Q}_s - \dot{Q}_t^{n-1}, \]

\[ \dot{Q}_s = (\alpha T) \exp[(T - 30)/16.7], \]

\[ \dot{Q}_t^n = \beta [M (\overline{\rho} + c^n) - M (\overline{\rho}^n)]. \]

Here \( u_a, v_a, p, c, \) and \( n \) are the zonal and meridional wind velocities, pressure, convergence anomalies, and iteration step, respectively. The linear frictional parameter \( \varepsilon \), gravity wave speed \( c_a \), SSTA heating parameter \( \alpha \), and the convective heating parameter \( \beta \) are given in ZC87. Function \( M(x) \) has the value

![Fig. 1. Composites of the observed SST (first row) and low-level wind (second row) anomalies during the mature phase (November–February mean) of the cold tongue (CT) El Niño (left column), warm pool (WP) El Niño (middle column), and La Niña (right column) events. The third row is for the wind responses to the observed SST anomalies in the original atmospheric model. The fourth row is for the differences between the model simulation and observation. For SST anomalies, the values with absolute values greater than 0.5 are shaded. For wind anomalies, the westerly anomalies for El Niño and easterly anomalies for La Niña are shaded.](image-url)
of $x$ for $x > 0$ and value 0 for $x \leq 0$. The readers are referred to ZC87 for detailed descriptions of the model and parameters. Note that $\overline{F}$ and $\overline{C}$ are the prescribed climatological monthly mean SST and low-level atmospheric convergence, respectively.

In this study, SST data are from the NOAA OISST (Reynolds et al. 2002), and convergence are calculated from 1000 to 925 hPa layer-mean winds of the NCEP-DOE Reanalysis II (Kanamitsu et al. 2002). As the Gill model treats the atmosphere of 400 m above the sea level as an entire layer, it is appropriate to use the abovementioned layer-mean winds as reference. In addition, the precipitation anomalies from GPCP v2.2 (Adler et al. 2003) are used for comparison of the atmospheric heating anomalies. Note that unit heating anomaly in the atmospheric model corresponds to 1.5 mm day$^{-1}$ precipitation anomalies (Zebiak 1990). For each variable, climatology is the monthly mean during 1981–2010 (1982–2010 for SST), and the anomalies are the departures from each monthly climatology. When evaluating the quality of the simulations, all the observed atmospheric anomalies are smoothed by a 3-month running mean filter to remove high-frequency signals.

As indicated by Zebiak (1990), the anomalous heating field produced by the ZC model is closely related to the SSTA in the tropical eastern Pacific, but the observed heating anomalies do not support it. That is, unrealistic convective heating anomalies occur over relatively low SST regions in the model. To eliminate this discrepancy, Kleeman (1991) proposed a threshold SST criterion of 28°C for deciding the occurrence of convection to reduce the anomalous heating release in the low SST regions (almost the CT region). However, Graham and Barnett (1987) and Waliser and Graham (1993) indicated that the relationship between convection and SST was a continuous function rather than a Heaviside function. Later, on the basis of Waliser’s observation, Wang and Li (1993) proposed a continuous function bridging convective heating and SST conditions in a simple coupled model. They concluded that the continuous heating function made the model perform better than Kleeman’s scheme in controlling convective heating release. In the ZC atmospheric model, we took a similar approach as Wang and Li (1993) to decide the anomalous convective heating under different types of SST and atmospheric circulation conditions. The heating efficiency factor $\delta$ is expressed as

$$
\delta = \begin{cases} 
1, & \text{if } T_s \geq 27.5^\circ C \text{ and } \text{Convergence} > 0, \\
(T_s - 25.5) \times 0.4 + 0.2, & \text{if } 25.5^\circ C \leq T_s < 27.5^\circ C \text{ and } \text{Convergence} > 0, \\
0.2, & \text{if } T_s < 25.5^\circ C \text{ or } \text{Convergence} < 0.
\end{cases}
$$

(6)

Here, $T_s$ is the total SST (mean SST and SST anomalies), and Convergence is the total low-level convergence. This function was designed on the following observational basis. As Waliser and Graham (1993) indicated, convection is well organized with a total SST of 27.5°C and is often associated with low-level convergence. Therefore, we assume that the anomalous heating release related to changes in SST and
low-level convergence under this condition has 100 % efficiency. For SST of 25.5–27.5°C, convection is not well developed. If the low-level circulation remains convergent, $\delta$ is assumed proportional to the total SST. For regions of low SST (< 25.5°C) or of low-level divergence, a constant heating coefficient of 20 % is assumed to preserve the anomalous heating that might not be caused by anomalous convection but is actually observed. Although the last setting is arbitrary, it does help simulate heating release better, as will be shown later. Under observed ENSO conditions, high coefficients appear over the WP and the Intertropical Convergence Zone (ITCZ), and low coefficients occupy the CT and off-equatorial regions in the Northern Hemisphere, as shown in Figs. 3a–c. When the model runs, this factor is calculated automatically at each time step using the model outputs.

After $\delta$ is added to modify the total heating anomalies, Eq. (3) becomes

$$
\varepsilon \rho \left( \frac{\partial \rho}{\partial t} \right) + c_d^2 \left[ \left( \frac{u_a^n}{\beta} \right)_x + \left( \frac{v_a^n}{\beta} \right)_y \right] = \delta (-\dot{\mathcal{Q}}_s - \dot{\mathcal{Q}}_{s-1}).
$$

By this modification, the model is expected to produce nearly unchanged heating anomalies over the central Pacific but reduce spurious heating anomalies in the eastern Pacific. Although the expression of $\dot{\mathcal{Q}}_s$ already ensures that anomalous heating is confined over the convergent region, this new $\delta$ makes it more strictly dependent on both the convergence and underlying SST.

After applying the heating coefficient to the heating equation, we further modify the atmospheric model by importing the effect of CMT into the zonal momentum equation. One candidate for simple formulation of CMT was proposed by Kang et al. (2010) in terms of mean vertical shear ($\overline{U}_z$) and anomalous low-level atmospheric convergence ($\varepsilon$). With this parameterization, the zonal momentum Eq. (1) is expressed as

$$
\varepsilon \rho \left( \frac{\partial \rho}{\partial t} \right) + c_d^2 \left[ \left( \frac{u_a^n}{\beta} \right)_x + \left( \frac{v_a^n}{\beta} \right)_y \right] = -\left( \frac{\partial p}{\partial y} \right)_x + A \overline{U}_z \varepsilon^n.
$$

Parameter $A$ is the efficiency of CMT for unit convergence anomaly. Kang et al. (2010) indicated that 0.003 was in a reasonable range of this parameter value for this type of shallow water model as they inferred from a general circulation model with CMT. $\overline{U}_z$ is defined as the vertical shear of the climatological monthly zonal wind between 200 and 925 hPa ($\overline{U}_{200} - \overline{U}_{925}$) over the ZC model domain. As Fig. 4 shows, over most of the tropical Pacific except for the equatorial regions west of the dateline, this is westerly shear irrespective of whether boreal winter or summer. Therefore, this term is expected to provide additional westerly (easterly) momentum to the lower level where convergence (divergence) anomalies are excited in the model. However, in conjunction with the CMT, easterly (westerly) anomalies in the western Pacific will be reduced because anomalous divergence (convergence) is generated and the CMT provides additional westerlies (easterlies) to cancel out the forced winds. We will show in Section 4 that this unwanted consequence is worth accepting for improving the simulation. Finally, Eqs. (8), (2), (7), (4), and (5) constitute our modified atmospheric model.
3. Optimal convective heating parameter used in the modified model

As shown in Section 2, thermal ($\delta$) and dynamical (CMT) modifications have been incorporated into the atmospheric model. However, a modification in the dynamics requires additional modifications, which can be dynamical or thermal or both, to rebalance the system. Additional dynamical modifications will unquestionably complicate the inherently simple framework of the model. Therefore, a reasonable choice is to modify the amount of anomalous heating for a given SST anomaly. In this section, we introduce the method to obtain the optimal convective heating parameter $\beta$ for our modified ZC atmospheric model.

The heating parameter $\beta$ is designed to vary in terms of its nondimensional reference value (0.75) in ZC87 with an increment of 0.05. The forcing is the observed monthly SST anomaly from 1982 to 2010, and the prescribed mean states are the same as those in Section 2. For each $\beta$, the corresponding zonal wind anomalies can be obtained from the new atmospheric model and compared with the observation. The criterion for evaluating the skill in simulating of zonal wind anomalies over the entire equatorial Pacific basin (120°E–80°W, 5°S–5°N) is straightforward: comparable amplitude, high correlation, and small root-mean-square (RMS) errors. Note that the response of modeled wind to heating is linear, and therefore changes in $\beta$ influence only the amplitudes rather than the spatial distributions of the heating and wind anomalies. The correlation between observation and each simulation of zonal wind anomalies will not vary with $\beta$, and thus the criterion of high correlation is not considered. Mathematically, the criterion can be expressed as three functions of $\beta$, as listed below (Eqs. 9–11).

\[
f(\beta) = |\text{stdv}(u^M_a) - \text{stdv}(u^O_a)|, \tag{9}
\]

\[
g(\beta) = \sqrt{\frac{\sum_{i=1}^{n} (u^M_{ai} - u^O_{ai})^2}{n}}, \tag{10}
\]

\[
F(\beta) = f(\beta) \times g(\beta). \tag{11}
\]

Here, stdv represents standard deviation and $u_a$ means the zonal winds anomalies. Superscripts M and O indicate the model simulation and observation, respectively. Subscript $i$ denotes each month in the period 1982–2010. Hence, the criterion becomes a mathematical question to determine for which $\beta$ function $F(\beta)$ reaches its minimum.

The dependences of the three functions on $\beta$ are shown in Fig. 5, in which the values for all three functions represent the area mean in the equatorial Pacific basin. The figure shows that function $f(\beta)$ reaches its minimum when $\beta$ is between 0.5 and 0.7 (Fig. 5a). Larger or smaller values of $\beta$ force the value of $f(\beta)$ away from its minimum, which means the amplitude of the simulated $u_a$ becomes less comparable with the observation. However, as shown in Fig. 5b, RMS errors increase with increasing $\beta$, indicating that larger values of $\beta$ cause larger simulated errors and vice versa. When both the comparable amplitude and small RMS errors are considered, the choice of the optimal value of $\beta$ becomes easier. As indicated in Fig. 5c, the value of $\beta = 0.58$ is the best mathematical choice. Therefore, in the modified model, the value of $\beta = 0.6$ is used.
4. Improvements in ENSO wind simulation

To examine the performance of the modified atmospheric model in simulating ENSO wind anomalies, the model is forced by mature SSTAs of CT-El Niño, WP-El Niño, and La Niña, as has been done using the original atmospheric model. These experiments are designed to highlight the improvements of the new model’s simulation by comparing it with original simulation and observation.

As a prerequisite of improvement, δ that is automatically calculated by the modified model must be mentioned first. For each ENSO condition, the pattern of the modeled δ resembles that of the observation, i.e., large values in the WP and ITCZ regions and small values in CT regions (Figs. 3d–f). Partly because of this factor, the simulated heating anomalies in the modified model (Figs. 6a, e, i) resemble those in the observation (Figs. 2a–c) better than those in the original model (Figs. 2d–f). In the modified model, the anomalous heating in the equatorial eastern Pacific during El Niño conditions is reduced by the modification of the heating coefficient (Figs. 6a, e); thus, it is easy to imagine that the projected heating anomalies onto the Kelvin waves are also reduced. Consequently, the easterly anomalies in the eastern Pacific are not as active as in the original model, which can be seen directly from the differences in the equatorial westerly anomalies between the two simulations (Figs. 6d, h). Moreover, the heating anomalies during the WP-El Niño event extend along a narrow zonal band in the ITCZ region (Fig. 6e), as the observation shows (Fig. 2b). Therefore, anomalous cross-equatorial southerlies are excited in the equatorial eastern Pacific (Fig. 6f). However, the original model underestimates these two features. For La Niña simulation, there are strong but overestimated negative heating anomalies confined in the ITCZ region, and unexpected weak positive heating anomalies appear near the eastern boundary in our proposed model (Fig. 6i). This pattern is in contrast with observation wherein strong negative heating anomalies exist in the central Pacific between 160°E and 160°W and weak negative heating anomalies exist along the ITCZ (Fig. 6j). Nevertheless, the differences between the simulated and observed wind anomalies in the La Niña condition are quite small (Fig. 6k) and are much smaller than the differences in the two El Niño conditions. Note that the wind differences between the two models for La Niña resemble an El Niño-like pattern of wind response (Fig. 6i), which means the new atmospheric model has weaker negative heating forcing. This reduction in anomalous heating is the exact purpose of the inclusion of δ. In summary, the simulations by the new model of the heating and wind anomalies associated with all ENSO conditions are improved with regard to reduced heating and wind anomalies in the eastern Pacific.

The improvements are more evident when comparing the horizontal distribution of the simulated zonal wind anomalies between the models and observation, as shown in Fig. 7. The original model overestimates the amplitude of the equatorial zonal wind anomalies (averaged between 5°S and 5°N), whereas the anomalous zonal winds from the modified model better fit the observation at most longitudes (Figs. 7a–c). The anomalous zonal winds in the eastern Pacific are reduced successfully; however,
the weakening of the easterlies in the western Pacific due to the additional westerlies from CMT during El Niño conditions is regarded as a minor and acceptable consequence of the modified model. The meridional scale of zonal wind anomalies in the central Pacific (160°E–150°W) tends to be broadened in the CT-El Niño simulation but slightly narrowed in both WP-El Niño and La Niña conditions (Figs. 7d–f). However, both models fail to reproduce the extensive southward shift of the anomalous westerlies in the central Pacific (Fig. 7d), which is a remarkable feature during the peak and decaying stages of the CT-El Niño events (Harrison and Vecchi 1999; Vecchi 2006; McGregor et al. 2013). Researches argue that such a shift emerged from the nonlinear interaction between the seasonal cycle and interannual ENSO variability (Stuecker et al. 2013), which involves seasonal migration of the WP to the Southern Hemisphere during the boreal winter and following spring (Stuecker et al. 2015; Xie and Yang 2014). If this nonlinear interaction is indeed the mechanism, then the most plausible reason for the failure of the models is that this nonlinear interaction is too weak. Because of weak presentation of this nonlinear mechanism, neither of the atmospheric models is able to reproduce the southward shift of the eastern Pacific ITCZ, therefore negative heating anomalies between 5° and 10°N during the CT-El Niño condition (Fig. 2a) are ignored by both models (Figs. 2d, 6a).

5. General capabilities of the modified model

Thus far, the improvements of the modified model in simulating wind anomalies under three categories of ENSO conditions have been introduced in detail. We now show some general features of the modified model in simulating wind response to SSTA forcing. The standard deviation and correlation maps of zonal wind anomalies in the two models versus the observation (Fig. 6). The third (fourth) row is for the wind differences between the modified model and observation (original model). The left, middle, and right columns are for the CT-El Niño, WP-El Niño, and La Niña episodes, respectively. The contour intervals of heating anomalies are 1 and the values with their absolute values greater than 1 are shaded. The westerly anomalies for El Niño and easterly anomalies for La Niña are shaded with heavy and light gray.

Fig. 6. The simulated total heating (first row) and winds anomalies (second row) in the modified atmospheric model as the responses to the observed SST anomalies in Fig. 1. The third (fourth) row is for the wind differences between the modified model and observation (original model). The left, middle, and right columns are for the CT-El Niño, WP-El Niño, and La Niña episodes, respectively. The contour intervals of heating anomalies are 1 and the values with their absolute values greater than 1 are shaded. The westerly anomalies for El Niño and easterly anomalies for La Niña are shaded with heavy and light gray.
The observed zonal wind anomaly shows a large number of distinctive features. Strong standard deviations are confined in the central and western Pacific (140°E–140°W, 10°S–10°N), whereas weak standard deviations appear in the eastern Pacific (Fig. 8a). The two simulations show similar features as the observation but with much narrower spatial extent in the central and western Pacific (Figs. 8b, d). In the eastern Pacific, both simulations show larger standard deviations than the observation. Nevertheless, we obtain confidence from the smaller local standard deviations in our modified model (Fig. 8b). Figures 8c and 8e show high correlations between the simulations and observation over the central and western Pacific, where the highest correlation appears between 160°E and 160°W. However, low and even negative correlations are still found in the eastern Pacific, indicating that the wind activities in this region are simulated poorly in both models. Even though we reduce the regional heating release in the modified model, the wind responses are opposite to that of the observation during ENSO (Fig. 5). Saji and Goswami (1996) reported that the winds in these regions are related to the surface pressure gradient rather than the convective heating, as described in the MG-type model. Furthermore, the wind response to anomalous pressure fields in the eastern Pacific is very weak in the ZC atmospheric model (Zebiak 1990). Undoubtedly, without an appropriate equation of the boundary layer dynamics neither model can reproduce such wind activities.

As it is commonly recognized that equatorial winds are of special importance in triggering air–sea interaction, we display the spatiotemporal evolution of the simulated equatorial zonal winds to obtain further insight into the capabilities of our modified model. In Fig. 9, we show the time evolution of the observed SST and zonal wind anomalies averaged over the central (Niño 4 region, 160°E–150°W, 5°S–5°N)}
and eastern (Niño 3 region, 150°W–90°W, 5°S–5°N) Pacific are shown in Fig. 10. The major fluctuations of zonal wind anomalies in the central Pacific from the two simulations are reproduced well, and the correlation coefficients between the simulations and observation are significantly high (Fig. 10a). We found that the original model overestimates the wind response in the central Pacific (Fig. 10a), particularly during ENSO events, as reflected by the large RMS error (1.94 m s\(^{-1}\)). Although the modified model show similar defects during the 1986–1988, 1992–1993, and 2009–2010 El Niño events, its performances during other El Niño events match the observation better and the RMS error drops down from 1.94 to 1.7 m s\(^{-1}\) (Fig. 10a). Moreover, the amplitude and duration of the easterly anomalies during most La Niña events are captured better and realistically by the modified model except during the weak 1984/85 La Niña event. As seen in Fig. 9e, there is almost no wind response to these negative SSTAs forcing in the modified model. In the eastern Pacific, the two models perform poorly in simulating wind response except for the passable simulations during the 1982/83 and 1997/98 strong CT-El Niño events (Fig. 10b). However, note that the simulated easterly anomalies associated with the continuously occurred WP-El Niño events between 2002 and 2007 are reduced in our modified model and the new simulation shows higher correlation and a smaller RMS error.

6. Model interpretation

Sections 4 and 5 demonstrate that our modified model shows substantial improvements in simulating ENSO winds. A question to be addressed is how this is achieved. In order to understand the main factors behind these improvements, we conduct further experiments to investigate the role of each modification.
The original model, as listed in Section 2, is marked as M1 and our new model (Eq. (12), also marked as M2 in the next context) is presented as follows. The modified terms are indicated in bold type.

\[
\begin{align*}
\varepsilon u_a^n - \beta_0 y v_a^n &= -\left(\frac{p^n}{\rho_0}\right)_x + A U_\infty c^n, \\
\varepsilon v_a^n + \beta_0 y u_a^n &= -\left(\frac{p^n}{\rho_0}\right)_y, \\
\varepsilon \left(\frac{p^n}{\rho_0} + c_a^2 \left[(u_a^n)_x + (v_a^n)_y\right]\right) &= \delta \left(-\tilde{Q}_s - \tilde{Q}_t^{n-1}\right), \\
\tilde{Q}_s &= (\alpha T) \exp[(T - 30)/16.7], \\
\tilde{Q}_t^n &= \beta_2 \left[ M (\sigma + c^n) - M (c^n) \right].
\end{align*}
\]

To bridge the solutions of M1 and M2, we introduce two middling models between them. For simplicity, all the models are forced by the mature SST anomalies of the CT-El Niño condition. The first middling model adopts the dynamical frame of ZC1 but uses the new convergence heating parameter \( \beta_2 \) from M2; we mark this model as M1.5a. Therefore, the only difference between M1.5a and M1 is the amount of the convergence heating anomalies. As shown in Fig. 11a, M1.5a produces less heating anomalies than M1, leading to weaker forced Rossby and Kelvin waves in M1.5a. Consequently, the associated westerly and easterly anomalies are reduced.

The other middling model (M1.5b) is designed on the basis of M1.5a but with \( \delta \) added in the heating term. Again, M1.5b is forced by the same SST anomalies, and the differences in the response of the winds and anomalous heating between M1.5b and M1.5a are displayed in Fig. 11b. With the inclusion of \( \delta \), heating anomalies in the southeastern tropical Pacific, where low heating coefficients exist, are reduced in M1.5b. Therefore, the differences between M1.5b and M1.5a present a La Niña-like wind response with anomalous easterlies (westerlies) in the central (eastern) Pacific (Fig. 11b). Note that the differences in wind anomalies are located mostly south of the equator because of the southerly location of the heating differences.
Models of M1.5a and M1.5b reduce the easterly anomalies in the eastern Pacific by suppressing anomalous heating release; however, simultaneous weakening in anomalous westerlies is ineluctable but unwanted. To compensate for these momentum losses, the dynamical effect of CMT is introduced into the zonal momentum equation of M1.5b, which means that M1.5b becomes M2. The differences show that the additional heating anomalies caused by CMT are very small, but the westerly anomalies supplied by CMT are relative large (Fig. 11c). Thus, the effect of CMT in this model is more dynamical than thermal. The compensation of the westerlies by CMT almost cancels out the losses of westerly anomalies due to heating suppression in the central Pacific (Fig. 11d), and the easterly differences in the central Pacific are due to the overestimation of wind response in M1, as has been mentioned before. Moreover, CMT enhances the westerly anomalies in the equatorial eastern Pacific (Fig. 11c). The direct differences between M2 and M1 (Fig. 11e) are similar to the sum of the difference between the two middling models (Fig. 11d), indicating that the thermal and dynamical modifications to the original atmospheric model are dynamically coherent.

We observed that the additional westerlies caused by CMT appear in the eastern equatorial Pacific (east of 160°W) in this simple model, whereas in the simulations of Kim et al. (2008) using a general circulation model, the westerly anomalies supplied by CMT locate in the central equatorial Pacific (west of 160°W, see their Fig. 11c). This difference arises mainly from the different locations of anomalous convergence/convection in the two models. In this simple model, anomalous convergence is generated in the eastern equatorial Pacific (figure not shown), following the pattern of SST anomalies, thus strong westerlies are induced by CMT in the eastern Pacific. But in the general circulation model used by Kim et al. (2008), anomalous convection is induced in the central equatorial Pacific, where the background SST is the highest, additional westerlies due to CMT are caused consequently in the central Pacific.

In fact, there is another indirect approach to evaluate the atmospheric response in M2. If we take the first-order derivative of heating to the low-level convergence, then we obtain $\frac{\partial \tilde{Q}_i}{\partial \tilde{c}^{n-1}} = 0$ and $\tilde{Q}_i^{n-1} = \frac{\partial \tilde{Q}_i^{n-1}}{\partial \tilde{c}^{n-1}} \cdot \tilde{c}^{n-1}$. Considering the small difference between $\tilde{c}^{n}$ and $\tilde{c}^{n-1}$ when the iteration approaches infinity, Eq. (7) is transformed to Eq. (15) as below. Together with the momentum budget equations, the differences between M2 and M1 can be
described approximately by the following system.

\[
\varepsilon \Delta u_a - \beta_0 y \Delta v_a = -\left(\Delta p / \rho_0\right)_x + A \mathcal{U}_z c, \quad (13)
\]

\[
\varepsilon \Delta v_a + \beta_0 y \Delta u_a = -\left(\Delta p / \rho_0\right)_y, \quad (14)
\]

\[
\varepsilon \left(\Delta p / \rho_0\right) + \left(c_a^2 + \delta \frac{\partial \hat{\Theta}}{\partial c}\right) \Delta c = 0. \quad (15)
\]

Here, the differences are described by the variables with \(\Delta\), and the anomalous convergence (c) in the CMT term \(A \mathcal{U}_z c\) can be approximated from the solution of M1. What is interesting is that these equations are analogous to the dynamical equations of the wind-driven upper-layer ocean model in the sense that the CMT term now acts like wind stress forcing in the ocean. It is well known that oceanic Rossby and Kevin waves forced by the westerlies tend to generate eastward currents both to the west and east of the forcing region. In switching this type of response to the atmospheric model, additional westerlies provided by CMT would generate atmospheric Rossby and Kevin waves in the atmosphere, and such waves would generate westerly winds in the regions to the west and east of the convection. As a result, CMT tends to enhance wind amplitude in the central Pacific where anomalous convergence is induced and reduce excessive easterly anomalies over the eastern Pacific. This is the principal reason why the inclusion of CMT can help maintain the realistic wind response to ENSO SSTA forcing when total heating anomalies are reduced by the inclusion of \(\delta\).

7. Summary and discussion

Aimed at reducing the eastern Pacific wind bias in the ZC atmospheric model, we have presented thermal and dynamical modifications and considerable improvements of ENSO winds simulation of the modified ZC model. The eastern Pacific “problem” of the original ZC atmospheric model is significantly reduced in the eastern Pacific.

Our simple modification to ZC atmospheric model keeps the same spirit of the simplicity of original formulation. We modified the simple convection parameterization in the original model by only adding a background-dependent efficiency factor into the convective heating, further adjusting one convective feedback parameter in the original model. This simple modification causes the convective heating to be dependent on the sum of mean SST and SST anomalies and the low-level convergence, which is consistent with the observed dependence of the convective activity on the observed SST and convergence distributions. Our simulated heating field shows better performance when compared with the observation than that of the original model simulation. Thereby, our wind simulation also improves.

Moreover, we add an ad hoc parameterization for the effect of CMT into the zonal momentum balance, and we further analyze the dynamic effect of the CMT. The dynamic effect of CMT-induced zonal momentum forcing can be understood similarly to the “wind forcing” to the ocean circulation in the shallow model context. With CMT parame-
terization, model simulation shows compensation to the over reduction in the central Pacific wind due to our modification in convective heating parameterization. Thus, these two modifications together show a significant improvement in simulated wind anomalies relative to that from the original ZC model and with the observation in terms of both amplitudes and spatial patterns.

These improvements are carried over to ZC coupled model as well when we add these modifications to coupled simulations. We conducted two 500-year simulations under the same background and the same initial condition using the original ZC model and the new version of the coupled model with our modifications to its atmospheric component. Two 50-year typical episodes from the evolutions of the equatorial SST and zonal wind anomalies from the simulations are shown in Fig. 12. In the simulation from new version of the coupled model, the spurious wind activities in the eastern Pacific are clearly much reduced, as it was the case in the uncoupled simulations shown in Fig. 9. However, the overall ENSO simulations such as ENSO pattern and irregularity in both simulations remain similar. The interdecadal variations in the amplitude of SST anomalies and the period of ENSO are discovered in both simulations. For instance, the ENSO activities during years 170–185 in the old model simulation (Fig. 12b) and during years 150–160 in the new model simulation (Fig. 12d) are less energetic and more irregular than those in other decades. Further detailed analysis, in particular the ability of the new model in capturing regimes of WP and CT ENSO and their different dynamics, will be reported in forth coming papers.

Fig. 12. Evolutions of the equatorial SST and zonal wind anomalies between year 150 and year 200 of the 500-year simulations when the original and modified atmospheric models are coupled with the same oceanic model from the ZC model. The contour intervals of SST and zonal wind anomalies are 1°C and 2 m s⁻¹, respectively.
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