Better Madden-Julian Oscillations with better physics: the impact of improved convection parameterizations

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Abstract

Two modifications are made to the deep convection parameterization in the NCAR Community Climate System Model, version 3 (CCSM3); a dilute plume approximation and an implementation of the convective momentum transports (CMT). These changes lead to significant improvement in the simulated Madden-Julian Oscillations (MJOs). With the dilute plume approximation, temperature and convective heating perturbations become more positively correlated. Consequently, more available potential energy is generated and the intraseasonal variability (ISV) becomes stronger. The organization of ISV is also improved, which is manifest in coherent structures between different MJO phases and an improved simulation of the eastward propagation of MJOs with a reasonable eastward speed. The improved propagation can be attributed to a better simulation of the background zonal winds due to the inclusion of CMT. We posit that the large-scale zonal winds are akin to a selective conveyor belt that facilitates the organization of ISVs into highly coherent structures, which are important features of observed MJOs. This study provides evidence for the interaction between the large-scale background state and the ISVs and concludes that it is necessary to consider MJOs in a multi-scale framework to enhance their understanding. The conclusions are supported by two supplementary experiments, which include the dilute plume approximation and CMT separately.
1. Introduction

Madden-Julian Oscillations (MJOs) are the dominant source of intraseasonal variance in the tropical atmosphere (Madden and Julian 1971, 1972, 1994, 2005). There have been numerous theories and model studies of MJOs to date, as summarized in Wang (2005) and Zhang (2005). Some theories have focused on the internal atmospheric dynamics, such as the wave-CISK (Lindzen 1974) and the ‘mobile wave-CISK’ (Lau and Peng 1987). Some emphasized the role of surface heat flux in the development of MJOs (Sobel et al. 2008, 2010), such as the ‘wind-induced surface heat exchange’ (WISHE) proposed by Emanuel (1987) and Neelin et al. (1987). Observations show that MJOs are also closely related to several phenomena occurring at different temporal and spatial scales, such as ENSO (Kessler and Kleeman 2000; Zavala-Garay et al. 2005), the North Atlantic Oscillation (Cassou 2008; Lin et al. 2009), and even the mean climate state (Sardeshmukh and Sura 2007). Therefore, it would appear necessary to study MJOs keeping in mind that their life cycle occurs in a multi-scale framework.

A discharge-recharge mechanism was proposed to emphasize the interactions between the large-scale circulation and small-scale convections (e.g., Blade and Hartmann 1993; Hu and Randall 1995; Kemball-Cook and Weare 2001). Biello and Majda (2005) established a multi-scale model which resolved both the upscale and downscale energy transfer associated with MJOs. In addition, Majda and Stechmann (2009) resolved convective momentum transport (CMT), which is critical for the interactions between the mesoscale and the planetary scale, in a simple dynamic model and reproduced major features of MJOs on multiple scales.

Climate model simulations of MJOs exhibit several substantial deficiencies. In particular, the eastward propagation of the convection over the warm pool is not well simulated (Slingo et al. 1996) and the eastward phase speed is not consistent with observations (Waliser 1999). Recently, Waliser et al. (2009) summarized the MJO diagnostics. The Level 2 diagnostics they proposed highlighted the organization (coherence) of MJOs between various variables and between different scales. They also emphasized that climate models have difficulty simulating this high degree of coherence. Therefore, it
is reasonable to argue that the observed MJOs are not only enhanced intraseasonal variabilities (ISVs) but, more importantly, are also well-organized ISVs. Kim et al. (2009) analyzed 8 model outputs following the standardized MJO diagnostics proposed in Waliser et al. (2009). They found that ECHAM4/OPYC displayed better skill in reproducing MJOs, which they argued was attributable to “a quite good mean state of precipitation and low-level wind”. In fact, possible improvement of MJO simulation due to an improvement in mean state has been proposed and tested in some other previous studies, such as Inness and Slingo (2003) and Sperber et al. (2005).

In this study, the major focus is on the influence of the background state on MJO simulations. Two modifications are made to the convection scheme in the Community Climate System Model, version 3 (CCSM3); implementing the dilute plume approximation and the CMT. As a result, the simulated MJOs become more energetic and the ISVs have more coherent structures with enhanced realism compared to observed MJO events. In Section 2, CCSM3 model configurations are described and differences in background states in the two model runs are highlighted. The improvements in the strength and the coherence of simulated MJOs are analyzed in Section 3 and 4, respectively. Conclusions are presented in Section 5.

2. Model description

CCSM3 is a state-of-the-art, global climate model which fully couples the atmosphere, ocean, land, and sea ice (Collins et al. 2006a, b). The perturbed deep convection model configurations are described in detail in Neale et al. (2008). Hence, only the relevant aspects are briefly mentioned below. The atmospheric component is based on the Community Atmosphere Model, version 3 (CAM3; Collins et al. 2006a) with 26 vertical levels and a horizontal resolution of 1.9° latitude × 2.5° longitude. There are 40 vertical levels in the ocean component with a nominal horizontal resolution of 1° × 1°. The parameterization of the deep convection basically follows Zhang-McFarlane scheme (Zhang and McFarlane 1995), but with two modifications. One is the dilute plume approximation (Raymond and
Blyth 1986, 1992), which relates the mixing between the reference parcel and the free troposphere to an assumed entrainment rate. The other one is the inclusion of CMT following Kershaw and Gregory (1997) which represents the compensation of convective momentum transport by atmospheric subsidence. More details about these two modifications can be found in Neale et al. (2008). Two fully coupled CCSM experiments were performed, one is the control run (referred to as C3OLD hereafter) without the above two changes and the other one includes the two modifications (referred to as C3NEW hereafter). Both experiments are conducted for 102 years and the last 20 years (Years 83 – Year 102) provide the period of focus for the daily output analysis. In the following, we will show that the two modifications to the Zhang-McFarlane scheme contribute in different ways to the improved MJO simulations. In order to further understand the separate role of the convection changes, two supplementary runs are performed; one including the dilute plume approximation (referred to as C3DPA hereafter) only and one including CMT (referred to as C3CMT hereafter) only. Both supplementary runs are conducted for 50 years and the last 10 years (Year 41 – Year 50) with daily outputs are used for analysis.

With the inclusion of CMT, the easterly bias of low-level zonal winds in the tropics is reduced, which has been reported in detail by Richter and Rasch (2008) and Neale et al. (2008). The improvement in the mean westerly winds can be seen in Fig. 1. Generally, the mean zonal winds at 850 hPa are quite similar in the two model runs (Fig. 1c and 1e). However, pronounced westerly winds from the central tropical Indian Ocean to the western Pacific Ocean (approximately from 50°E to 180°E; Fig. 1c) are reproduced, although they are still a little weaker than reality as represented by the NCEP reanalysis (Kalnay et al. 1996; Fig. 1a) over the maritime continent (Fig. 1d). The climatological winds in C3CMT (not shown) are very similar to those in C3NEW (Fig. 1c). In contrast, the westerly winds can barely be found in the above region in C3OLD (Fig. 1e and 1f). Except for the background zonal winds, other background fields are very similar in the two model runs. The mean geopotential height and surface temperature in C3NEW, which are representatives of the dynamic and
thermodynamic fields, respectively, are shown in Fig. 2a and 2c. The corresponding differences between C3NEW and C3OLD are shown in Fig. 2b and 2d. For both quantities, relatively large differences (but still very small compared with the mean state) are found over the central tropical Pacific Ocean (around 200°E), northern Pacific Ocean, and the southern Indian Ocean. Over the Indo-Pacific warm pool, which is the region we are interested in, the differences are even smaller. Latent heat flux in C3NEW, which can play a critical role in tropical ISVs (e.g., Sobel et al. 2008, 2010), is shown in Fig. 2e and the differences are shown in Fig. 2f. From the tropical Indian Ocean to the western tropical Pacific Ocean, the difference is generally between ±5 W m⁻², which is less than 5% of the mean value in this region. In addition, the vertically averaged moist static energy (MSE, \( h = C_p T + L q + g z \), where \( C_p \) is the specific heat of the air, \( T \) is the air temperature, \( q \) is the specific humidity, \( g \) is the gravitational acceleration, and \( z \) is the geopotential height) in C3NEW is shown in Fig. 2g, along with the difference between the two model runs (Fig. 2h). Since MSE is an indicator of the stability of the air column (see further discussion below), the high similarity between the two model runs indicates that the stabilities of the background state in the two model runs have no distinct differences. Note however that these stabilities are reached after adjustments in the coupled systems and the state variables such as winds and precipitation do display differences in spatial and temporal scales. In all, the only significant difference in the background state between the two model runs resides in the enhanced westerly winds over the tropical Indian Ocean and the western Pacific Ocean in C3NEW. Except the mean westerly winds, all other background variables are very similar to each other in the two experiments. In this study, we will show that a better organization of simulated MJOs in C3NEW is closely related to the improved background westerly winds.

3. Strength of the intraseasonal variabilities

Following the method proposed in Slingo et al. (1999), the strength of the simulated MJOs is represented by the variance of the intraseasonal zonal winds at 200 hPa over the tropical Indian Ocean
(averaged from 10°S to 10°N and from 50°E to 100°E and passed through the 101-day running mean, Fig. 3). In this study, all intraseasonal variables are obtained with a band-pass filtering from 20 to 100 days. With the same method, Zhang and Mu (2005; their Fig. 2) showed that applying a modified Zhang-McFarlane scheme for deep convection could considerably increase the strength of ISVs in CCSM3. In C3OLD (which uses the Zhang-McFarlane scheme with no modifications) there are indeed energetic ISVs such as the peaks seen at the end of Years 91 and 95 (Fig. 3), which reaffirms that adopting Zhang-McFarlane scheme can indeed produce strong ISVs in a model. Nevertheless, the ISVs in C3NEW are in general stronger than those in C3OLD, e.g., in Years 84, 88, 96, 99, and 101. The stronger ISVs in C3NEW can also be clearly seen in other variables, such as the intraseasonal zonal winds at 850 hPa and the outgoing longwave radiation (OLR; see Neale et al. 2008). Note that the impacts of model differences in monsoons, ENSO, and their interactions on the ISVs are beyond the scope of this work and are not considered here.

At 850 hPa, the standard deviations (STDs) of the intraseasonal zonal winds in C3NEW (Fig. 4c) are similar to those in NCEP reanalysis (Fig. 4a). Generally, the differences between C3NEW and NCEP are smaller than 0.5 m s\(^{-1}\) over the tropical Indian Ocean and the western Pacific Ocean (Fig. 4b). However in C3OLD, the ISVs in this region are not distinctly different from other tropical regions (such as the central Pacific Ocean and the Atlantic Ocean; Fig. 4e). As clearly shown in Fig. 4d and 4f, the major enhancement in the intraseasonal zonal winds in C3NEW is over the Indo-Pacific warm pool. Therefore, with the modifications to the deep convection parameterization (Neale et al. 2008), the tropical ISVs in C3NEW become generally more energetic than those in C3OLD.

The mean vertically averaged deep convective heating (product of the Zhang-McFarlane scheme) in the two model runs are compared in Fig. 5a and b. Over the tropical Pacific Ocean, the double ITCZ problem still exists in C3NEW, although there is moderate improvement in contrast with C3OLD (note the negative anomalies around 10°S in the central Pacific in Fig. 5b). It was argued in Neale et al. (2008) that the current changes in convection scheme are not very helpful for removing the southern
Pacific convergence zone (SPCZ) bias. Another notable feature in Fig. 5b is the negative values from
the tropical Indian Ocean to the western tropical Pacific Ocean, the region in which the ISVs are
actually stronger in C3NEW than those in C3OLD (Fig. 4d). In the intraseasonal band (20-100 days),
the STDs of convective heating in C3NEW (Fig. 5c) are also smaller than those in C3OLD (Fig. 5d).
In addition, precipitation in the two model runs displays no pronounced differences (not shown).
However, as emphasized in Emanuel et al. (1994), the convective heating is not the key factor for ISVs;
instead, the correlation between convective heating and the temperature perturbation is the key. The
thermodynamic equation can be written as
\[
\frac{\partial T'}{\partial t} - w' S_p = \frac{J'}{C_p} \tag{1}
\]
where \( T \) is the air temperature, \( w \) is the vertical velocity, \( S_p \) is the static stability parameter, \( J \) is
convective heating which is the major component of the total external heating, \( C_p \) is the heat capacity
of the air, and the prime denotes the intraseasonal component (between 20 days and 100 days) of the
variables. Multiplying \( T' \) on both sides of Eq. (1), one has
\[
\frac{\partial}{\partial t} \left( \frac{T'^2}{2} \right) - T' w' S_p = \frac{T' J'}{C_p} \tag{2}
\]
where \( T'^2 / 2 \) is proportional to the potential energy in the intraseasonal band, \( T' J' \) is critical to the
buildup of available potential energy during convection and \( T' w' \) determines the energy conversion
from potential energy to kinetic energy. The two terms \( T' J' \) and \( T' w' \) are distinctly different between
the two model runs. Similar to the method used above in Fig. 3 (Slingo et al. 1999), we calculate the
mean variance of \( J' \) and \( T' J' \) (vertically averaged, then averaged from 10°S to 10°N and from 50°E to
100°E and passed through a 101-day running mean). As shown in Fig. 6a, the variance of convective
heating is virtually indistinguishable between the two runs, which is consistent with Fig. 5. In contrast,
the variance of \( T' J' \) in C3NEW is clearly larger than that in C3OLD, which explains the more
energetic ISVs in the former. The model performance supports the principle of the quasi-equilibrium
theory for convection (Emanuel et al. 1994). Due to the dilute plume approximation, the environmental
air is entrained into the cloud at all levels, not only at the cloud top (Raymond and Blyth 1986). As a result, the phase relations between the convective heating and the temperature perturbation are modified. Although the convective heating is generally a little weaker because of the entrainment of dry air in C3NEW than it is in C3OLD (Figs. 5 and 6), both the buildup of potential energy ($T'J'$) and the release to kinetic energy ($T'w'$) become larger in C3NEW. Therefore, the ISVs in C3NEW are stronger than those in C3OLD. This conclusion is also supported by the comparisons between C3DPA and C3CMT in Fig. 3b. Stronger ISVs in C3DPA compared to C3CMT points to the fact that the reinforcement of ISVs is attributable to the dilute plume approximation.

### 4 Organization of the intraseasonal variabilities

#### 4.1 Composite MJO phases

Differences between the two model runs reside in the propagating features of the ISVs. The ISVs in C3NEW are not only more energetic, but also better organized in comparison to reality. Thus, they resemble the observationally-inferred MJOs, much more so than the ISVs in C3OLD.

Observed MJOs are composed of successive suppressed and active phases during the eastward propagation (Zhang 2005). Thus, with the EOF analysis, the first two EOF modes are in quadrature, which is represented with the significant cross-correlation at a time lag of 10-15 days between the principal components (PCs) of the first two EOFs. This relation is well captured in C3NEW in the intraseasonal zonal winds at both 200 hPa and 850 hPa, and for the intraseasonal OLR anomalies, while it is not well captured in C3OLD (Fig. 7). For comparison with reality, the cross-correlation between RMM1 and RMM2, which were defined in Wheeler and Hendon (2004), is superimposed in the middle panel in Fig. 7. Better coherence between the first two EOF modes of the intraseasonal zonal winds and OLR indicates a better eastward propagation of the ISVs during MJOs, which is a major challenge for MJO simulations. The ISVs in C3OLD, although they can be energetic, do not have such coherent eastward propagation. In addition, the first two EOF modes in C3NEW explain
about 14% of the total variance of the intraseasonal zonal winds at 850 hPa and about 16% of the total variance at 200 hPa, which are comparable with the observations (13% – 16%). As summarized in Waliser et al. (2009), “many climate (model) simulations produce leading EOFs for convective fields that explain relatively small amounts of the variance compared to observations”, since the ISVs in many models were likely to be strong enough but not organized well enough. Cross-correlations between the first two EOFs for C3CMT and C3DPA are also shown in Fig. 7. Although the ISVs in C3CMT are weaker than the counterparts in C3DPA (Fig. 3b), the quadrature relation is better captured in C3CMT, which implies that the CMT is more conducive to the organization of the ISVs. However, the first two EOFs in C3CMT only explain about 6% and 5% respectively of the total variance in the intraseasonal band, because the organized ISVs are weak due to the absence of the dilute plume approximation.

Following Wheeler and Hendon (2004), the MJO phases can be defined with the first two PCs of the multivariate EOF analysis (based on OLR, zonal winds at 850 hPa and 200 hPa) and the phase diagram is shown in Fig. 8. When $PC_1^2 + PC_2^2$ is larger than 2, the ISVs are regarded as strong.Canonically, an MJO event originates from the western Indian Ocean and diminishes over the eastern Pacific Ocean. Thus, the line rotates counterclockwise in the phase diagram. Six MJO events are defined based on the strength of the ISVs (represented with $PC_1^2 + PC_2^2$) and the eastward propagation (the phase diagram shown in Fig. 8a), which are listed in Table 1. For comparison with reality, the phase diagram of 4 observed MJO events are shown in Fig. 8b. The number of MJO events appears to be fewer than reality in typical years. One important reason is that we intend to guarantee that all selected MJO events can travel continuously around the globe from the western Indian Ocean back to the western hemisphere since we want to focus on the eastward propagation of the ISVs in following discussion. Thus, the selected MJO events are required to show smooth counterclockwise rotation in the phase diagram as shown in Fig. 8. This criterion is much stricter than only requiring large amplitudes (like $PC_1^2 + PC_2^2 > 2$). The composite intraseasonal OLR anomalies in the 8 MJO
phases are shown in Fig. 9. In Phase 1, there is almost no depression in OLR over the tropics. In Phases 2 and 3 when the convection center is supposed to be over the Indian Ocean, negative OLR anomalies are discernible from the western tropical Indian Ocean to the central Indian Ocean. Meanwhile the OLR depression becomes stronger in Phase 3 than it is in Phase 2. In Phase 4 and 5 when MJOs reach a mature stage over the maritime continent, the negative OLR anomalies reach their maxima. Then from Phase 6 to Phase 8, negative OLR anomalies move eastward from the maritime continent to the western hemisphere while they weaken along the way. The whole process of the simulated composite MJO event in C3NEW is very similar to observations. Since the MJO phases can only be well defined in C3NEW (ISVs in C3CMT are weak; ISVs in C3OLD and C3DPA hardly propagate and the variances explained by the first two EOF modes are relatively small), all composite quantities shown below are calculated with the C3NEW outputs.

The difference in the intraseasonal MSE between the surface and the mid-troposphere is a good index for the atmospheric instability (Kemball-Cook and Weare 2001); this instability usually being a necessary ingredient for the onset of MJOs. Defining $\Delta MSE$ as the intraseasonal MSE at 950 hPa minus the intraseasonal MSE at 500 hPa, positive $\Delta MSE$ implies an atmosphere unstable to saturated moist convection, while negative $\Delta MSE$ implies a stable atmosphere. Figure 10 shows the composite $\Delta MSE$ in the 8 MJO phases during the 6 MJO events (Table 1). In Phase 3, positive $\Delta MSE$ begins to occur over the Indian Ocean and the maritime continent, indicating deep convective instability in these regions. In Phase 4, $\Delta MSE$ reaches a maximum over the maritime continent, where the MJOs also become mature (see the OLR anomalies in Phase 4 in Fig. 9). In Phase 5 and Phase 6, positive $\Delta MSE$ turns weak and the MJOs enter the decaying phases. In Phases 7 and 8, there is no clear positive $\Delta MSE$ in the deep tropics, thus the MJO convective signals become weaker in the western hemisphere. One may see that $\Delta MSE$ increases down the SPCZ in Phases 7 and 8 which leads to a split in negative OLR anomalies during these phases. The MJO signal bifurcates too strongly here (this happens weakly in observations), which may be due to cold SST biases in the central and eastern Pacific Ocean. Actually,
ΔMSE is a practical and convenient measure of the convectively available potential energy (CAPE).

The composite intraseasonal CAPE anomalies also display a similar evolution (not shown) as shown in Fig. 10. Maloney (2009) concluded that surface heat flux and horizontal advection are the dominant terms in the intraseasonal MSE budget.

The composite intraseasonal latent heat fluxes during the 6 MJO phases are shown in Fig. 11. In Phase 1, one can see positive surface heat flux in the western tropical Indian Ocean, which is enhanced in Phase 2 and extends to the central and eastern Indian Ocean. In Phase 3, the surface heat fluxes are a maximum over the eastern Indian Ocean and the maritime continent. Then in Phase 4, positive surface heat fluxes remain over the maritime continent and the southeastern Indian Ocean, but become weaker than they are in Phase 3. The maximum in surface heat fluxes (in Phase 3) is generally one phase (about 5 – 7 days) earlier than the mature stage of MJOs (Phase 4), which is roughly consistent with the observations (Sperber 2003). From Phase 5 to Phase 7, positive surface heat fluxes move eastward to the western and central Pacific Ocean. The composite convergence of MSE ($-\vec{u} \cdot \nabla h$; Maloney 2009) in the 8 phases is shown in Fig. 12. The patterns are somewhat noisy. But some features can still be clearly identified if we focus on the tropical region. In Phases 1 and 2, MSE converges over the western Indian Ocean. In Phase 3, the horizontal transport of MSE increases over the whole tropical Indian Ocean and the maritime continent and then becomes weaker in Phase 4. Synthesizing the information from Fig. 10 to Fig. 12, it can be concluded that over the Indian Ocean and the maritime continent (Phase 3 and Phase 4), the air column over the eastern Indian Ocean and the maritime continent becomes unstable due to the surface heat flux and horizontal transport. As a result, the increased potential energy is released and transferred to kinetic energy and the reinforced intraseasonal winds occur in these regions.

The above analyses on MSE are consistent with the analysis of convective heating in Section 3, both of which explain how the ISVs become stronger in C3NEW. However, the ISVs in C3NEW propagate eastward (Fig. 7) and resemble realistic MJO events, while the ISVs in C3OLD barely
propagate, although they are enhanced occasionally (Fig. 3). Naturally, the question is what leads to the eastward propagation of the enhanced ISVs in C3NEW and more specifically what determines the eastward propagation speed of MJOs? Since the distinct difference between C3OLD and C3NEW model simulations resides in the simulated background low-level westerly winds from the tropical Indian Ocean to the western Pacific Ocean (Fig. 1), the better organization of ISVs in C3NEW should be attributable to the improved westerly winds in C3NEW.

4.2 Eastward propagation of MJOs

Several mechanisms have been proposed to explain and simulate the slow movement of the MJO events, which is much slower than the free planetary waves. A friction-convergence mechanism was invoked to account for the retardation in the eastward propagation of MJO events (Wang and Rui 1990; Salby et al. 1994). However, the drag coefficients used in the friction-convergence theory were deemed to be unrealistically large, such that the actual role of friction was also debatable (Sperber et al. 1997; Moskowitz and Bretherton 2000). In a numerical model study, Chao and Chen (2001) showed that friction may not be as important as claimed for the low-level convergence. Instead, friction only dissipates the energy of MJOs. Studies have also shown that the enhanced surface heat flux could determine the speed of the eastward-propagating ISVs, as hypothesized in the WISHE theory (Emanuel 1987; Neelin et al. 1987). In practice, a weak, rather than a strong evaporation usually leads the convection (Woolnough et al. 2001). Furthermore, it is nontrivial to explain why most MJO events “select” to travel around a specific speed of 5 m s\(^{-1}\), rather than a random speed over a wide range. With our model simulations and NCEP reanalysis, we will show that the large-scale zonal winds are likely to play a role in pacing the eastward speed of MJOs.

As described above, the generation of available potential energy and kinetic energy during convection is determined by \( T'J' \) and \( T'w' \). Actually we can treat \( T'J' \) and \( T'w' \) in a similar way when assessing the circumstances under which these two terms have significantly non-zero values.
According to the weak temperature gradient approximation (Sobel et al. 2001; Bretherton and Sobel 2002) which was thoroughly tested for applications in the tropical ISVs (Bretherton and Sobel 2003), variation of $T'$ with longitude is small and the principal thermal-dynamical balance is $-w'S_p = J'$, where $S_p$ is the static stability parameter. Since $S_p$ is a slowly-varying quantity, $w'$ and $J'$ have similar structures. The weak temperature gradient and similar structures between $w'$ and $J'$ are also reproduced in C3NEW. As shown in Fig. 13, the variation patterns of $J'$ and $w'$ are similar and the variation of $T'$ along a latitude is quite small compared with the variation of $J'$ and $w'$. In order to quantify the above argument, along every latitude, we calculate the ratio $\text{STD}(X_{lat})/\text{mean}(X_{lat})$, where $X_{lat}$ represents the STD of $T'$, $J'$, and $w'$ along latitudes shown in Fig. 13(a, c, e). The ratios are shown on the right column of Fig. 13. For $T'$, the zonal variance is less than 10% of the zonal mean, while for $J'$ and $w'$, the zonal variance is about 20% – 60% of the zonal mean. Therefore, the zonal oscillatory parts of $T'.J'$ and $T'.w'$ are determined by $J'$ and $w'$, respectively. $J'$ and $w'$ are composed of oscillations with various wavenumbers and frequencies (Wheeler and Kiladis 1999) and each component can be written in a wavelike form, i.e. $A(k)e^{it\psi}$, where $\psi = kU - \omega(k)$ and $U = x/t$ is the mean zonal current. Thus, $T'.J'$ can be expressed as the sum of the waves with all wavenumbers (the same argument can also be applied to $T'.w'$),

$$T'.J' = T' \int_0^\infty A(k)e^{it\psi} dk$$ \hspace{1cm} (3).

Then integration by parts yields

$$T'.J' = T' \int_0^\infty A(k)e^{it\psi} \frac{d}{dk} = \frac{T' A(k)e^{it\psi}}{it} \bigg|_0^\infty - T' \int_0^\infty e^{it\psi} \frac{d}{dk} \left( \frac{A}{d\psi / dk} \right)$$ \hspace{1cm} (4).

According to the method of stationary phase (see Pedlosky 2003 for details), the term $\frac{T' A(k)e^{it\psi}}{it} \bigg|_0^\infty$ decreases with $1/t$ and approaches zero after a considerable period of time (when $t$ is large). The last term in Eq. (4) also decreases at least as fast as $1/t$, unless $e^{it\psi}$ does not oscillate too rapidly at large $t$. 

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Therefore, $T'J'$ has a significantly non-zero value only when $\psi$ does not change at large $t$, in other words, at the stationary point of $\psi$ with respect to $k$. Thus, $\partial \psi / \partial k = 0$ is a necessary condition and the group velocity of perturbations is $c_g = \partial \omega / \partial k = U$. This relation implies that the propagation speed of the perturbations, which can last for a long time, is not determined by the meso-scale variability itself. Instead, it is selected by the background zonal flow. Note that we do not argue that the ISVs are simply advected by the mean flow. Actually, the fluctuations caused by an individual convective event can have considerably different propagation speed from the background state (Hendon and Liebmann 1994; Wheeler and Kiladis 1999). These fluctuations can be very energetic in the spatio-temporal scales of tropical convection (on the order of days and hundreds of kilometers), but tend to dissipate on much larger spatio-temporal scales (e.g., the scale of MJOs, on the order of tens of days and thousands of kilometers), because they have random phases which are prone to cancel each other (the physical essence of the method of stationary phase). Only the fluctuations which have a coherent relation with the background state can last for a long time and propagate over large distances. This process is analogous to a selective conveyor belt (the background state), which selectively picks only the suitable perturbations and discards (dissipates) the unfitting ones, organizing the former into an energetic event which is much larger in space and longer in time than any individual perturbation. As a result, one can observe the well-organized MJO events, which originate from the un-organized deep convection but have much larger space scales and much longer time scales than any individual deep convective cell.

The above arguments are supported by evidence from both our model simulations and the NCEP reanalysis (Kalnay et al. 1996). Intraseasonal OLR anomalies during two simulated MJO events are shown in Fig. 14 (a and b). The ISVs propagate eastward at a speed of about 4 m s$^{-1}$, as indicated with the gray dashed lines. Contours of the background zonal winds at 4 m s$^{-1}$ at 850 hPa are superimposed. The eastward propagation speed of ISVs is consistent with the background zonal wind speed. As shown in Fig. 3, there are also pronounced ISVs in C3OLD, but they do not propagate to the east (see the flat lag-correlation coefficients in Fig. 7). Intraseasonal OLR anomalies from the end of Year 95 to
the beginning of Year 96 in C3OLD are shown in Fig. 14c. Strong negative OLR anomalies in the central Indian Ocean (around 100°E) are obvious and they are comparable with those in C3NEW in strength. However, the background zonal winds are very weak during this period of time and the negative OLR anomalies in C3OLD do not propagate to the east. Therefore, it is reasonable to assume that the distinctive difference in the eastward propagation of ISVs in the two model runs is attributable to the difference in the simulations of the large-scale zonal winds. This hypothesis is also supported when one looks at OLR observations and NCEP reanalysis. Figure 15 shows the intraseasonal OLR anomalies (obtained from NOAA polar-orbiting series of satellites, Liebmann and Smith 1996) during four MJO events, superimposed on the zonal background velocities (from NCEP reanalysis). The propagation speeds of the intraseasonal OLR anomalies, which are marked with gray lines, range from 4 to 4.7 m s\(^{-1}\). The speeds are consistent with the zonal background speeds. Thus, both observations and our simulations confirm that the eastward propagation speeds of MJOs are consistent with the background zonal wind speed.

The above analyses can help us understand why the eastward propagation of ISVs in C3NEW is better than those in C3OLD. As shown in Fig. 1, the background westerly winds are pronounced and around 5 m s\(^{-1}\) in C3NEW from the tropical Indian Ocean to the western Indian Ocean. As a result, in C3NEW and C3CMT, the ISVs are better organized by the large-scale flow and propagate eastward, which is represented with the cross-correlation between the first two EOF PCs shown in Fig. 7 (especially for the intraseasonal zonal winds at 850 hPa and for intraseasonal OLR anomalies). In contrast, in C3OLD and C3DPA, there are almost no background low-level westerly winds over the Indo-Pacific warm pool. Therefore, although the ISVs can be enhanced by the Zhang-McFarlane deep convection parameterization scheme (Zhang and Mu 2005) and the dilute plume approximation, their eastward propagation tends to be rather weak. Instead, the ISVs are mostly generated locally and dissipated also locally (Fig. 14c). As a result, there is no significant cross-correlation between the first two EOF PCs (Fig. 7), which indicates a lack of eastward propagation.
5. Conclusions

Two modifications are made to the convection scheme in the CCSM3 model, which leads to a better simulation of MJOs. The climatologies of most variables are not significantly changed by the modifications to the Zhang-McFarlane scheme, except for the mean westerly winds from the Indian Ocean to the western Pacific Ocean. In a coupled system, the post-adjustment states may well appear proximate but the adjustment process itself may produce more distinguishable responses, as in the strength and propagation of MJOs for the cases under consideration here. Inclusion of dilute plume approximation improves the correlation between $J'$ and $T'$ which is critical for the buildup of the available potential energy. Although the convective heating is indistinguishable between C3NEW and C3OLD, $T'\cdot J'$ is much larger in C3NEW. Since $w'$ has a similar structure to $J'$, $T'\cdot w'$ is also larger in C3NEW, which makes the energy conversion from potential energy to kinetic energy more efficient. As a result, ISVs in C3NEW are enhanced in strength. In the additional experiment C3DPA, it is also shown that inclusion of the dilute plume approximation helps to reinforce ISVs. With the inclusion of CMT, the low-level background zonal winds over the Indo-Pacific warm pool are reinforced and the easterly wind bias in this region is largely removed. The improved zonal winds tend to pace the eastward propagation of the enhanced ISVs, acting like a selective conveyor belt. In the experiment C3CMT, although the ISVs are not strong enough, the eastward propagation of ISVs is also discernable due to the improved westerly winds.

As articulated in Waliser et al. (2009), the high degree of coherence is a very important feature of MJOs. By comparing the model runs with CCSM3 with and without the dilute plume approximation and the CMT, we show that a better simulation of the background westerly winds in the tropical Indian Ocean and the western tropical Pacific Ocean facilitates a better organization of the simulated MJOs. The quadrature relation between the first two EOF modes is well captured in C3NEW and the eastward propagation is similar to the realistic MJOs. Both with the NCEP reanalysis and the model outputs, it is
shown that the eastward propagation speed of the intraseasonal signals is tuned by the background zonal currents. The ISVs themselves, which can be regarded as stochastic perturbations to the background state (e.g., Newman et al. 2009), are not as well organized. Nevertheless, only the ISVs that are coherent with the background state can accumulate energy and propagate over a relatively long distance. Thus, the ISVs are selected (not just advected) and organized by the background state. Notice that the background zonal winds are not independent of the mesoscale convection, since CMT represents the connection between convection and the background state. Thus, the MJO speed is essentially the result of an interaction between the mesoscale convection and the background circulation. With an aqua-planet model, Maloney et al. (2010) demonstrated that the synchronization between peak moistening and total westerly winds at 850 hPa determines the eastward propagation speed of the simulated MJOs, which was argued to also highlight the role of the coupling between convection and background winds in selecting the MJO speed.

This conclusion confirms the importance of examining MJOs in a multi-scale framework. A better simulation of the background state is helpful to achieve a better simulation of MJOs. In addition to the deep convection parameterization being regarded at present to be a major requirement for improved MJO simulations, this also calls for more attention to better representation and parameterization of the interactions between the large-scale circulation and the mesoscale ISVs. In particular, the hypothesis that the propagation speed of MJOs is closely related to the mean westerly winds is also consistent with the improvement of the eastward propagation of the simulated MJOs even when they are approximately resolved in a model. The background states are also affected by changes in climate modes such as monsoons and ENSO. While that is consistent with the multi-scale framework we are arguing for, the details of those interactions are not addressed here other than that they impact the background westerlies discussed here. Further details of the monsoon-ENSO-MJO interactions will be reported elsewhere.
Acknowledgments:

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Captions

Table 1 MJO events defined with the C3NEW outputs.

Figure 1 Mean zonal winds at 850 hPa in the NCEP reanalysis (a), C3NEW (c), and C3OLD (e). Differences between C3NEW and C3OLD, between NCEP and C3NEW, between NCEP and C3OLD are shown in the right column. The unit is m s\(^{-1}\).

Figure 2 20-year mean geopotential height, surface temperature, latent heat flux, and moisture static energy (MSE) in C3NEW (left column) and the corresponding differences between C3NEW and C3OLD (right column; regions with positive values are shaded).

Figure 3 Variance of the intraseasonal zonal winds at 200 hPa averaged from 10°S to 10°N and from 50°E to 100°E and then passing a 101-day running mean. (a) Comparison between C3NEW and C3OLD. (b) Comparison between C3CMT and C3DPA.

Figure 4 STDs of intraseasonal zonal winds at 850 hPa in NCEP (a), C3NEW (c), and in C3OLD (e). The differences between the STDs in NCEP and C3NEW are shown in (b). The differences and the ratios between the STDs in C3NEW and C3OLD are shown in (c) and (d), respectively.

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Figure 7 Cross-correlations between PC1 and PC2 of the intraseasonal zonal winds at 200 hPa, at 850 hPa, and intraseasonal OLR anomalies. Cross-correlations between RMM1 and RMM2 defined in Wheeler and Hendon (2004) are superimposed in the middle panel with a dot-dash line.
Figure 8 (a) Phase diagram in terms of the first two PCs of the EOF analysis for 6 MJO events in C3NEW listed in Table 1. Every curve rotates anti-clockwise. The dash cycle is the unit cycle. (b) is the same as (a) but for 4 observed MJO events which are used again in Fig. 15.

Figure 9 Composite intraseasonal OLR anomalies during 6 MJO events in C3NEW (Table 1). The unit is W m\(^{-2}\).

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### Table and Figures

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</tr>
<tr>
<td>6</td>
<td>Day 65/Yr 84</td>
<td>Day 126/Yr 84</td>
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</tbody>
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