Upscale Impact of Mesoscale Convective Systems and its Parameterization in an Idealized GCM for a MJO Analog above the Equator

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ABSTRACT

The Madden-Julian oscillation (MJO) typically contains several superclusters and numerous embedded mesoscale convective systems (MCSs). It is hypothesized here that the poorly simulated MJOs in current coarse resolution global climate models (GCMs) is related to the inadequate treatment of unresolved MCSs. A satisfactory understanding of the upscale impact of MCSs on the MJO is lacking, so its parameterization should provide the missing collective effects of MCSs. A simple two-dimensional multicloud model is used as an idealized GCM with clear deficiencies. Eddy transfer of momentum and temperature by the MCSs, predicted by the mesoscale equatorial synoptic dynamics (MESD) model, is added to this idealized GCM. The upscale impact of westward-moving MCSs promotes eastward propagation of the MJO analog, consistent with the theoretical prediction of the MESD model. Furthermore, the upscale impact of upshear-moving MCSs significantly intensifies the westerly wind burst, due to two-way feedback between easterly vertical shear and eddy momentum transfer with low-level eastward momentum forcing. Finally, a basic parameterization of upscale impact of upshear-moving MCSs modulated by deep heating excess and vertical shear strength is provided as a new parameterization. This significantly improves key features of the MJO analog in the idealized GCM with clear deficiencies. A three-way interaction mechanism between the MJO analog, parameterized upscale impact of MCSs and background vertical shear is identified.
1. Introduction

The MJO is the dominant component of tropical intraseasonal variability (Zhang 2005) and dramatically impacts local weather through extreme rainfall and mid-latitude atmospheric conditions through tropical-extratropical teleconnection (Zhang 2013; Stan et al. 2017; Henderson et al. 2017). Tropical convection associated with the MJO is hierarchically organized across multiple spatial and temporal scales. The MJO typically contains multiple eastward- and westward-moving superclusters of cloudiness (Nakazawa 1988; Chen et al. 1996) with numerous embedded MCSs (Houze 2004) and cumulus clouds on smaller scales. As the major rainfall producer in the tropics, MCSs contribute up to 50% of the rainfall in most tropical regions (Tao and Moncrieff 2009).

Although the effects of large-scale atmospheric conditions on the modulation of MCSs have been well documented in observations (Lin and Johnson 1996; Chen et al. 1996; LeMone et al. 1998), a satisfactory understanding of the collective effects of MCSs on the momentum and heat budgets of the MJO is still lacking.

It is hypothesized that the poorly simulated MJOs in current coarse resolution GCMs are related to the inadequate treatment of MCSs and their upscale impact. Typical behavior of the poorly simulated MJOs in the GCMs includes impersistent eastward propagation, unrealistic planetary/intraseasonal variability in precipitation and winds, and upright vertical structure with negligible westerly wind burst (WWB) (Jiang et al. 2015). In contrast, global cloud-resolving simulations that resolve MCSs successfully capture some key features of the MJOs (Grabowski 2003; Miura et al. 2007), which motivated the development of the superparameterization method based on two-dimensional cloud resolving models (CRMs) (Grabowski 2001, 2004; Randall et al. 2003; Majda 2007a) and a sparse space-time technique (Xing et al. 2009). Nevertheless, the computational cost to explicitly resolve MCSs is too expensive to be practical. An alternative way to
address this issue is to develop new parameterizations for coarse-resolution GCMs that capture
the upscale impact of unresolved MCSs on the MJO.

Several studies have assessed the upscale impact of MCSs based on observational measurement,
reanalysis dataset and cloud-resolving simulations, most of which focus on convective momentum
transfer (CMT) (Moncrieff 1981; LeMone 1983; Moncrieff 1992; LeMone and Moncrieff
1994). Convective-scale CMT by unorganized convection normally has frictional effects that re-
duce large-scale vertical shear (Zhang and McFarlane 1995). In contrast, mesoscale CMT by
organized convection over hundred kilometers in horizontal scale can have countergradient mo-
mentum transport that enhances the large-scale vertical shear (Moncrieff 1981, 1992). Tung and
Yanai (2002a) concluded that CMT is, on the average, downgradient over the western Pacific
warm pool but upgradient during the westerly wind phase of the MJO (Tung and Yanai 2002b).
Oh et al. (2015) found that the subgrid-scale and mesoscale CMT associated with the MJO has
a distinctive three-layer vertical structure. Grabowski and Moncrieff (2001) demonstrated that
CMT from westward-moving MCSs embedded in the eastward-moving convective envelope pro-
motes the large-scale organization of convection. Inspired by multi-scale organization and the
observed statistical self-similarity in tropical convection, Majda (2007b) systematically derived
multi-scale asymptotic models that describe scale interactions among clusters, superclusters and
intraseasonal oscillations and highlight the crucial role of eddy transfer of momentum and tem-
perature. Brenowitz et al. (2018) concluded that mesoscale CMT dominates the total vertical flux
feedback on planetary-scale kinetic energy budget, providing new mechanisms for the planetary-
scale organization of convection.

From a theoretical perspective, several modeling studies have been undertaken to better under-
stand the upscale impact of MCSs on the large-scale organization of tropical convection. Majda
and Stechmann (2009) utilized a simple dynamic model with features of CMT from convectively
coupled gravity waves and their interactions with large-scale mean flow. Khouider et al. (2012) demonstrated that in the active region of the MJO with WWB, CMT from both convectively coupled Kelvin waves (CCKWs) and MCSs plays a significant role in accelerating the low-level westerly winds. The three-dimensional mesoscale equatorial synoptic dynamic (MESD) model, originally derived by Majda (2007b), was used as a multi-scale framework to assess the upscale impact of MCSs on eastward-moving CCKWs (Yang and Majda 2018b) and westward-moving 2-day waves (Yang and Majda 2018a). Explicit expressions for eddy transfer of momentum and temperature obtained from the MESD model provide an essential basis for parameterization of upscale impact of MCSs here. Moncrieff et al. (2017) introduced the multi-scale coherent structure parameterization (MCSP) that achieved significant improvement in tropical precipitation patterns and precipitation variability in a GCM.

The goals of this paper include the following three aspects: first, use a simple multicloud model for the MJO analog and intraseasonal variability above the equator to mimic the typical behavior of GCMs with clear deficiencies. Secondly, assess the upscale impact of MCSs on key features of the MJO analog, including persistent propagation of a two-scale structure, realistic planetary/intraseasonal variability in precipitation and winds, and a significant WWB. Thirdly, introduce a basic parameterization of upscale impact of MCSs and test its effects in the idealized GCM to deal with deficiencies.

In general, the multicloud models describe three dominant cloud types (congestus, deep, stratiform) by using the 1st- and 2nd-baroclinic modes and build the life-cycle of these cloud types into the convective heating closure through the usage of a switch function for mid-latitude dryness (Khouider and Majda 2006c, 2007, 2006a). The deterministic version of the multicloud models successfully captures characteristic features of CCEWs (Khouider and Majda 2008b, 2006b, 2008a) and the diurnal cycle (Frenkel et al. 2011a,b, 2013), and the stochastic version captures the
MJO (Khouider et al. 2010; Deng et al. 2015; Goswami et al. 2017) when coupled to the GCM. In this paper, we use a deterministic two-dimensional multicloud model for the MJO analog and intraseasonal variability above the equator (Majda et al. 2007; Harlim and Majda 2013). By reducing the magnitude of both congestus and stratiform heating, this model mimics the typical behavior of GCMs with clear deficiencies, where both convection types are underestimated (Seo and Wang 2010; Del Genio et al. 2012; Lappen and Schumacher 2012; Del Genio et al. 2015). In order to introduce the upscale impact of MCSs, we use explicit expressions for the eddy transfer of momentum and temperature as theoretically predicted by the MESD model (Yang and Majda 2018b).

The upscale impact of MCSs on the MJO analog is assessed through comparison experiments with/without adding extra eddy transfer of momentum and temperature. To mimic the scenario that MCSs are prominent in the active convection region of the MJO (Khouider et al. 2012), the modulation effects of deep heating excess on eddy transfer of momentum and temperature are considered. The results show that the upscale impact of westward-moving MCSs promotes the eastward propagation of the MJO analog, consistent with the theoretical prediction by the MESD model (Yang and Majda 2018b). The modulation effects of vertical shear are considered to mimic the observation that MCSs typically move towards the convection center (Lin and Johnson 1996; Chen et al. 1996; Moncrieff and Klinker 1997; Yanai et al. 2000; Houze Jr et al. 2000). The results show that the upscale impact of upshear-moving MCSs leads to a significant WWB in the middle and west of the MJO analog, due to the positive feedback between large-scale easterly vertical shear and embedded eddy momentum transfer with low-level eastward momentum forcing. Finally, we provide a basic parameterization of upscale impact of upshear-moving MCSs, where modulation effects of deep heating effects and vertical shear strength are linearly combined. Significant improvement is achieved upon adding this parameterization into the idealized GCM that
has clear deficiencies. A further simulation illustrates a three-way interaction mechanism between the MJO analog, parameterization of upscale impact of MCSs and background mean flow over a long time scale. The resulting oscillatory background mean flow on time scales of several years resembles the QBO-like oscillation as typically seen in cloud resolving simulations (Held et al. 1993; Nishimoto et al. 2016) and simplified GCMs (Horinouchi and Yoden 1998).

The results of this paper are presented as follows. Section 2 summarizes the governing equations and properties of the two-dimensional multicloud model, including the realistic MJO analog above the equator and the idealized GCM that has clear deficiencies. Section 3 discusses the effects of eddy transfer of momentum and temperature from MCSs on the MJO analog. Section 4 provides a basic parameterization of upscale impact of upshear-moving MCSs under the modulating effects of deep heating excess and vertical shear strength and tests its effects in the idealized GCM that has clear deficiencies. The paper concludes with discussion in Section 5.

2. An Idealized GCM for a MJO Analog and Intraseasonal Variability above the Equator

In this section, we briefly review the governing equations of the multicloud model and the convective heating closure. The simple two-dimensional multicloud model used here (Majda et al. 2007; Harlim and Majda 2013) captures the MJO analog and intraseasonal variability above the equator. The goals of this section are to reproduce (1) realistic MJO analog above the equator as a proxy for the observations and (2) a simulation with reduced congestus and stratiform heating as an idealized GCM having clear deficiencies.

a. Governing equations and multicloud model parameterization

Multicloud models describe the life-cycle of the three main cloud types (congestus, deep and stratiform) (Johnson et al. 1999) and incorporate it in the convective heating closure by using
a switch function for mid-tropospheric dryness. In detail, shallow congestus convection is first
initialized with low-level heating (upper-level cooling), moistening the lower troposphere and pre-
conditioning the deep convection. Then the deep convection warms the whole troposphere by
bringing extreme rainfall, followed by stratiform convection with upper-level heating by latent
heat release and low-level cooling due to rain evaporation (Khouider and Majda 2008b).

The governing equations and multicloud convective parameterization in dimensionless units are
listed in Table.1 and Table.2, and all relevant parameters and constant are listed in Table.3. All
physical variables are nondimensionalized by synoptic scaling. Specifically, the speed of the first-
baroclinic dry Kelvin waves \(c = N \frac{H_T}{\pi} = 50ms^{-1}\) is the horizontal velocity scale, the equatorial
Rossby deformation radius \(L = \sqrt{\beta c} = 1500km\) is the length scale, \(T = \frac{L}{c} = 8hrs\) is the time
scale, \(\tilde{\alpha} = \frac{H_T \Theta_0}{\pi g} N^2 = 15K\) is the temperature scale, and \(\tilde{\beta} = 45Kday^{-1}\) is the heating scale.

Consistent with the 1st-baroclinic deep heating and the 2nd-baroclinic consgestus/stratiform
heating, both momentum and temperature variables in the free troposphere are truncated to the
1st- and 2nd-baroclinic modes by using the following Galerkin projection,

\[
f = f_1 \left[ \sqrt{2} \cos(z) \right] + f_2 \left[ \sqrt{2} \cos(2z) \right], \quad f \in \{u, p, F_u\} \tag{1}
\]

\[
g = g_1 \left[ \sqrt{2} \sin(z) \right] + g_2 \left[ 2\sqrt{2} \sin(2z) \right], \quad g \in \{\theta, S^\theta, F^\theta\} \tag{2}
\]

where \(u\) is zonal velocity, \(p\) is pressure perturbation, \(F_u\) is eddy momentum transfer, \(\theta\) is potential
temperature anomaly, \(S^\theta\) is heating, and \(F^\theta\) is eddy heat transfer. As shown by Table.1, the
1st- and 2nd-baroclinic momentum is forced by linear momentum damping mimicking boundary-
layer turbulent drag \(-\frac{C_{nu}}{H_b} u_j\) and Rayleigh friction \(-\frac{1}{\nu_R} u_j\) as well as eddy momentum transfer \(F^u_j\).

The 1st-baroclinic potential temperature is driven by the deep heating \(P\) and the 2nd-baroclinic
potential temperature is driven by congestus and stratiform heating \(-H_s + H_c\), both of which are
further forced by radiative cooling \(-Q_{R,j}^0 - \frac{1}{\sigma_D} \theta_j\) and eddy heat transfer \(F^\theta\). These dynamical
fields are coupled to a column-integrated moisture perturbation (Khouider and Majda 2006b), where both linear and nonlinear moisture advection terms are retained and precipitation $-\frac{2\sqrt{7}}{\pi}P$ and downdrafts $\frac{D}{Hf}$ are added at the right-hand-side as moisture sink and source, respectively.

The boundary-layer equivalent potential temperature equation shows that surface-level evaporation $\dot{E}_{hb}$ warms and moistens the boundary layer while the downdrafts $\frac{D}{h_b}$ have the opposite effects.

Both congestus heating $H_c$ and stratiform heating $H_s$ are governed by linear relaxation equations, where congestus heating is triggered in the regions with cold and dry mid-troposphere before deep heating, and stratiform heating lags the deep heating. A switch function for mid-troposphere dryness $\Lambda$ is defined in Table.2. The multicloud heating closure is completed by introducing deep heating $P$, downdrafts $D$ and evaporation $E$.

All physical variables are imposed on the domain of the tropical belt, $0 \leq x < 40,000$ km, with periodic boundary conditions in the zonal direction. The governing equations as shown in Table.1 and Table.2 are solved numerically by spatially discretizing the solutions at equal-spacing grids and then temporally integrating time steps by using the 4th-order Runge-Kutta scheme. The horizontal resolution is 100 km and each time step is 4.5 min, close to the typical coarse-resolution GCMs. The moisture equation with nonlinear advection terms is solved by pseudo-spectra methods. To stabilize the numerical scheme and eliminate grid-scale numerical instability, a fourth-order hyper-diffusion term is also added to all prognostic equations, $-\nu f_{xxxx}$, where the dimensionless value of $\nu$ is chosen as $2 \times 10^{-5}$ based on the trial-and-error strategy.

A convenient way to discuss linear convective instability in the multicloud model is to consider the radiative-convective equilibrium (RCE) state. Specifically, we consider a state where zonal velocity, $u = 0$, and potential temperature and moisture perturbation vanish in both troposphere and boundary layer, $\theta_j = 0, \theta_{eb} = 0$ and $q = 0$. The actual value of the other variables at the RCE state is included in Table.4. To trigger unstable moist modes, a random field of moisture in a very
weak magnitude \(10^{-5}\) in dimensionless units) is added to the initial conditions. All solutions presented in this paper are obtained in the equilibrium state after the simulations are integrated over a long period (4000 days in Section 2 and 3, 7000 days in Section 4).

\[b. \text{Realistic MJO analog and intraseasonal variability above the equator}\]

Here we first implement the 2D multicloud model with all default parameter values as Majda et al. (2007). In these default parameters, the congestus and stratiform adjustment coefficients are 
\[\alpha_c = 0.5\text{ and } \alpha_s = 0.25,\text{ respectively, and the background moisture stratification } \bar{Q} \text{ is 1.0. Although the typical value of } \bar{Q} \text{ in other studies based on observation has smaller value 0.9, a larger value of } \bar{Q} \text{ is chosen as Majda et al. (2007) to increase convective instability and intensify precipitation.}\]

We run the simulation for 3000 days and use the last 1000-day output in the equilibrium state for interpretation purposes. Since the model output in the default parameter regime share several realistic features as observation, we regard it as a realistic MJO analog and intraseasonal variability above the equator. We will use this as a proxy for observations.

Fig.1a shows the Hovmoller diagram for precipitation during the last 200 days. The solutions are characterized by a two-scale structure with eastward-moving planetary-scale envelopes and numerous embedded westward-moving synoptic-scale disturbances. The planetary-scale envelopes have wavenumber 2 and a period of 40 days that propagate eastward at a speed of \(6.17 \text{ ms}^{-1}\). Inside these planetary-scale envelopes, several synoptic-scale disturbances are embedded and propagate westward at an even slower speed, resembling the observed westward-moving superclusters in the active phases of MJO over the West Pacific, such as the 2-day waves (Chen et al. 1996). Fig.1b and Fig.1c show the power spectra of precipitation and zonal velocity based on model output during the last 1000 days. As for precipitation, its eastward-moving component has a dominant peak in wavenumber 2 and period of 30 days. The power spectra of zonal velocity is similar but confined
to a smaller wavenumber and longer period, consistent with the typical observation that the dynamical circulation usually has larger spatial scales than the heating that drive it. Fig.1d-e show the zonal and vertical profiles of the composite planetary-scale envelopes in the moving reference frame. As shown by panel (d), the precipitation peak is led by both column-integrated moisture and boundary-layer equivalent potential temperature and followed by stratiform heating. This is consistent with the intuition that a moist free troposphere and boundary layer tends to precondition deep convection while stratiform convection typically forms subsequent to deep convection in the form of anvil clouds. Panel (e) shows the vertical cross-sections of zonal velocity and potential temperature anomalies in the free troposphere. Both fields are characterized by a front-to-rear tilt, akin to the observed features of the MJO. The surface-level westerlies resemble the WWB of the observed MJO.

Key features of a realistic MJO analog include the following three aspects: First, two-scale structure with eastward-moving planetary-scale envelope and embedded westward-moving synoptic-scale disturbances. Secondly, power spectra of precipitation and zonal velocity with dominant peaks at wavenumber 1-3 and period of 30-90 days in eastward-moving components and wide bands of power signals for westward-moving components at wavenumber 5-15 and period less than 30 days. Thirdly, front-to-rear tilts in zonal velocity and potential temperature with the WWB in the middle and west of the planetary-scale envelope. In the remaining experiments, we will mainly focus on these three key features of the MJO analog.

c. An idealized GCM with clear deficiencies

Sensitivity experiments (not shown) with this model show that the solutions are quite sensitive to several key parameters, such as stratiform heating adjustment coefficient $\alpha_s$, congestus heating adjustment coefficient $\alpha_c$ and background moisture stratification $\tilde{Q}$. There is no guarantee that
these key parameters will have optimal values in some physically motivated application, resulting in significant bias and poor behavior. In order to mimic the typical behavior of GCMs with clear deficiencies, we reduce the heat adjustment coefficients for congestus $\alpha_c$ and stratiform convection $\alpha_s$ to half as shown by Table.3. Meanwhile, the background moisture stratification $\tilde{Q}$ is increased to 1.03 from 1.00 for relatively stronger convective instability.

Fig.2a shows Hovmoller diagrams for precipitation after the system reaches the equilibrium state. The planetary-scale envelopes have wavenumber 4, which is somewhat shorter than the observed MJO wavelength in the wavenumber 1-3 range. Meanwhile, these planetary-scale envelope propagates eastward at a speed of $2.4 \text{ m s}^{-1}$, much slower than the typical propagation speed of the observed MJO ($5 \text{ m s}^{-1}$). The maximum magnitude of precipitation is about $8 \text{ K day}^{-1}$, much weaker than that in Fig.1. Fig.2b and Fig.2c show log-scale power spectra of precipitation and zonal velocity based on the last 1000-day model output. Notably, these power spectra peaks are quite symmetric about the axis wavenumber zero, both of which are featured by the planetary-scale (near wavenumber 4) and intraseasonal (near 40 days) variability. Such eastward/westward symmetry stems from the mixture of both eastward- and westward-moving MJO analog. In fact, present-day GCMs suffer a similar bias in that the power spectra of westward-moving planetary-scale precipitation is as significant as the eastward-moving counterpart.

3. Upscale Impact of Mesoscale Convective Systems on the MJO Analog above the Equator

In this section, we assess the upscale impact of MCSs on the MJO analog through comparison experiments with/without eddy transfer of momentum and temperature from mesoscale fluctuations. Specifically, we use the idealized GCM with clear deficiencies in Fig.2 as the control run. In order to introduce the upscale impact of MCSs, we use the explicit expressions for eddy transfer of momentum and temperature obtained from theoretical predictions of the MESD model (Yang
and Majda 2018b) and consider upscale impact of MCSs that propagate either slowly \((5 \text{ ms}^{-1})\) or rapidly \((20 \text{ ms}^{-1})\), either upshear or downshear, modulated by the deep heating excess and vertical shear strength. Due to the zonal symmetry property of the model in Table.1, we only need consider the case with westward-moving MCSs because the opposite case can be inferred through mirror reflection.

**a. Eddy transfer of momentum and temperature predicted by the MESD model**

In general, multi-scale models are derived by following multi-scale asymptotic methods (Majda and Klein 2003; Majda 2007b) and have been applied to study multi-scale interactions of tropical convection such as the upscale impact of synoptic-scale fluctuations on the MJO (Majda and Biello 2004; Biello and Majda 2005, 2006), the intraseasonal impact of the diurnal cycle on the MJO (Yang and Majda 2014; Majda and Yang 2016) and ITCZ breakdown (Yang et al. 2017). In particular, the Majda (2007b) MESD model has been used to assess upscale impact of embedded MCSs on eastward-moving CCKWs (Yang and Majda 2017, 2018b) and westward-moving 2-day waves (Yang and Majda 2018b). In those studies, mesoscale heating is prescribed by phase-lagged first- and second-baroclinic modes to mimic the front-to-rear tilt structure in observations (Houze 2004),

\[
s_\theta = c_0 \left[ \sin (kx' - \omega \tau) \sin (z) + \alpha \sin (kx' - \omega \tau + \phi_0) \sin (2z) \right]
\]  

(3)

where \(x'\) points to the propagation direction of mesoscale heating. \(c_0\) is magnitude coefficient. \(k\) and \(\omega\) are wavenumber and frequency respectively. Here \(\alpha\) is the relative strength coefficient of the second-baroclinic mode, and \(\phi_0\) the phase lag. The MESD model theoretically predicts the
explicit expressions for eddy transfer of momentum and temperature,

\[ F^\mu = \kappa^\mu \left[ -\frac{3}{2} \cos(z) + \frac{3}{2} \cos(3z) \right] \cos(\gamma), \quad \kappa^\mu = \frac{c_0^2 \sin(\phi_0) \alpha k^3}{2(\omega^2 - k^2)(4 \omega^2 - k^2)} \]  

(4)

\[ F^\theta = \kappa^\theta \left[ \frac{3}{2} \sin(z) - \frac{9}{2} \sin(3z) \right], \quad \kappa^\theta = \frac{c_0^2 \sin(\phi_0) \alpha k^3 c}{2(\omega^2 - k^2)(4 \omega^2 - k^2)} \]  

(5)

where \( \gamma \) is the tilt angle between propagation direction of mesoscale heating and zonal direction.

In the following experiments, for simplification, \( F^\mu \) and \( F^\theta \) are further truncated by only retaining the dominant first-baroclinic mode.

Fig. 3 shows vertical profiles of mesoscale fluctuations and eddy transfer of momentum and temperature. In particular, the red curves in panels (c,d) show the corresponding eddy transfer of momentum and temperature for eastward-propagating mesoscale systems. When the mesoscale systems propagate westward, the sign of eddy momentum transfer is simply reversed, while that of eddy heat transfer stays the same. Here we truncate the vertical profiles of eddy transfer of momentum and temperature by only retaining the first-baroclinic mode. That is, the eddy momentum transfer has eastward (westward) momentum forcing in the lower (upper) troposphere, which reach maximum strength at the surface (top) of the domain. The eddy heat transfer cools throughout the troposphere, and its maximum strength is in the middle troposphere.

It is straightforward to induce that the ratio between \( F^\theta \) and \( F^\mu \) in dimensionless units is determined by propagation speed of the mesoscale heating,

\[ \frac{\kappa^\theta}{\kappa^\mu} = c \]  

(6)

where \( c \) is the dimensionless value (dimensional value divided by 50 ms\(^{-1}\)) of propagation speed of the mesoscale heating. In the following simulations, we do not need to specify exact values of parameters in the expressions of \( \kappa^\mu, \kappa^\theta \), but just specify the value of \( \kappa^\mu \). Then the value of \( \kappa^\theta \) is inferred by Eq.6, when the propagation speed of the mesoscale heating \( c \) is also specified.
b. Eddy transfer of momentum and temperature modulated by deep heating excess

Here we consider the scenario when the eddy transfer of momentum and temperature in the first baroclinic mode is modulated by the maximum allowable deep heating excess $P_0$ as follows,

$$F^u = \kappa^u \frac{P_0^+}{\bar{Q}} \left[ -\frac{3}{2} \cos(z) \right]$$ (7)

$$F^\theta = \kappa^\theta \frac{P_0^+}{\bar{Q}} \left[ \frac{3}{2} \sin(z) \right]$$ (8)

where $P_0 = \frac{1}{\tau_{\text{conv}}} (a_1 \theta_{eb} + a_2 q - a_0 (\theta_1 + \gamma_2 \theta_2))$ is the anomaly component of the maximum allowable deep heating. $\bar{Q}$ is the corresponding RCE value. The value of the expression $f^+$ stays the same if $f$ is positive and zero if it is negative. The closure for $P_0$ is a combination of the Betts-Miller relaxation-type parameterization and convective available potential energy (CAPE) parameterization. Physically, the maximum allowable deep heating excess $P_0$ resembles the effect of CAPE in modulating MCSs and the resulting CMT (Moncrieff 2004). Majda and Stechmann (2008) developed a stochastic parameterization for CMT, whose strength is modulated by the square of the maximum allowable deep heating.

Three cases are compared here with/without $F^u$ and $F^\theta$ modulated by the effects of $P_0$. The first case is the same as the control run in Fig.2. The second and third cases consider the eddy transfer of momentum and temperature from MCSs that propagate at a slow ($5 \text{ ms}^{-1}$) and fast speed ($20 \text{ ms}^{-1}$). The magnitude coefficient for eddy momentum transfer $\kappa^u$ is fixed at 0.0032. The difference between the second and third cases lies in the stronger magnitude of $F^\theta$ in the case of fast propagation.

Fig.4 shows the Hovmoller diagrams for precipitation. Compared with the control run in panel a, the cases with eddy terms in panels b and c show an apparent two-scale structure, where planetary-scale envelopes propagate eastward and embedded synoptic-scale disturbances propagate westward. In panel b, the maximum magnitude of precipitation reaches 28 K/day. Such intense
precipitation and the promoted eastward-moving planetary-scale envelope by westward-moving
MCSs is consistent with the previous results of Yang and Majda (2018b). In panel c, the maxi-
mum magnitude of precipitation is reduced to 12 K/day and convection is largely suppressed due
to the extra cooling from eddy heat transfer, again consistent with the previous results of Yang and
Majda (2018b). This extra cooling reduces low-level moisture convergence, resulting in weaker
growth rate of the unstable modes.

Fig. 5 shows the power spectra of precipitation and zonal velocity. Compared with the control
run in panel a, both cases show clear east/west contrast in the power spectra, similar to the real-
istic MJO analog as shown in Fig. 1. In the slowly propagating MCSs case, the power spectra of
precipitation in panel c are characterized by three discrete spectra peaks for the eastward-moving
components and three bands of power spectra of westward-moving components. In particular,
the peak for eastward-moving planetary-scale envelope has wavenumber 3 and period around 50
days. The power spectra of zonal velocity in panel d is quite similar to that in panel c, indicating
a close correlation between convection and the large-scale circulation. As for the case with faster
propagating MCSs in panels e-f, the associated power spectra of precipitation are dominated by a
planetary-scale peak for eastward-moving component and a band of power spectra for westward-
moving component.

Fig. 6 shows the vertical cross-sections of composite planetary-scale envelopes in the moving
reference frame. The vertical structure of zonal velocity and potential temperature anomalies fea-
tures a significant front-to-rear tilt, consistent with the built-in life cycle transition from congestus
to deep to stratiform convection. In panel a, the maximum magnitude of zonal velocity of about 2
ms$^{-1}$ is reached at the top of the domain. In the lower troposphere, the wind convergence is mostly
in phase with the maximum precipitation with westerlies to the west and easterlies to the east. The
WWB is negligible here. The maximum magnitude of both positive and negative potential tem-
perature are both reached in the upper troposphere. In contrast, both the maximum magnitude of zonal velocity, potential temperature anomalies and precipitation anomalies in panel b are much weaker than those in panel a, indicating suppressed convection due to eddy heat transfer.

c. Eddy transfer of momentum and temperature modulated by vertical shear

Here we consider the scenario when eddy transfer of momentum and temperature is modulated by the strength of vertical shear $\triangle U$ as follows,  

$$F^u = \kappa^u \frac{\triangle U}{U_{ref}} \left[ -\frac{3}{2} \cos (z) \right]$$  

$$F^\theta = \kappa^\theta \frac{\triangle U}{U_{ref}} \left[ \frac{3}{2} \sin (z) \right]$$

where $U_{ref} = 50 ms^{-1}$ and the strength of vertical shear is defined as follows,

$$U^{u}_{max} = \max_{\pi/2 \leq z \leq \pi} \{u\}; U^{u}_{min} = \min_{\pi/2 \leq z \leq \pi} \{u\}$$  

$$U^{l}_{max} = \max_{0 \leq z \leq \pi/2} \{u\}; U^{l}_{min} = \min_{0 \leq z \leq \pi/2} \{u\}$$

$$\triangle U \equiv \max \left\{ \left| U^{u}_{max} - U^{l}_{min} \right|, \left| U^{u}_{min} - U^{l}_{max} \right| \right\}$$

Fig.7a explains the definition of vertical shear strength $U^{l}_{max}$, which basically calculates the maximum possible easterly and westerly shear between the upper and lower troposphere and selects the larger one. Fig.7b describes the scenarios when the MCSs propagate upshear (along the opposite direction of vertical shear) and downshear (along the same direction of vertical shear).

Four cases are compared with/without $F^u$ and $F^\theta$ modulated by the effects of $\triangle U$. Besides the first cases from the control run in Fig.2, the remaining three cases consider the eddy transfer of momentum and temperature from MCSs that propagate westward, upshear and downshear at a slow speed (5 $ms^{-1}$). Correspondingly, the magnitude coefficient of eddy momentum transfer $\kappa^u$
is 0.0024, 0.0030, 0.0030, respectively. The choice of a smaller value of $\kappa''$ in the second case is to obtain a more realistic precipitation intensity.

Fig.8 shows the Hovmoller diagrams for precipitation. Compared with the control run in panel a, the maximum magnitude of precipitation in both panels b and c is intensified, while that in panel d is weakened. Specifically, the maximum magnitude of precipitation in panel b reaches 25 $K/day$, which is consistent with the previous result that westward-moving MCSs favors the eastward propagation of convection (Yang and Majda 2018b). The patterns of spatio-temporal variability of precipitation in panels b is featured by the two-scale structure with eastward-moving planetary-scale envelopes at wavenumber 3 and embedded westward-moving synoptic-scale disturbances at shorter wavelength. Compared with the realistic MJO analog in Fig.1, the solutions exhibit more intermittency in terms of variable precipitation intensity and spatio-temporal patterns. In panel c, the maximum precipitation in panel c also intensifies to 19 K/day, mainly due to the positive feedback between vertical shear and eddy momentum transfer. The precipitation anomalies are dominated by both eastward- and westward-moving planetary-scale envelopes and exhibit no clear east/west contrast. Based on a similar argument, the precipitation in panel d is mainly reduced due to the negative feedback between vertical shear and eddy momentum transfer from downshear-moving MCSs.

Fig.9 shows the power spectra of precipitation and zonal velocity for these three cases. Compared with the symmetric power spectra in the control run, the cases in panels c-d are characterized by significant zonal asymmetry. Specifically, the eastward-moving components are dominated by a continuous band of power spectra along the non-dispersive line across the equator, which extends from wavenumber 3 to 10 and period from 15 days to 50 days. In this case, such continuous power spectra reflect the intermittent nature of both precipitation and zonal velocity fields. For the
case in panels e-f, the power spectra of both precipitation and zonal velocity exhibits large zonal
symmetry, indicating the prevalence of both eastward- and westward-moving MJO analogs.

Fig.10 shows vertical cross-sections of zonal velocity and potential temperature anomalies. No-
tably, WWB at the surface does not occur in panel a, while it has a much stronger magnitude in
panel b. In the case with eddy terms from westward-moving MCSs, the eddy momentum trans-
fer tends to induce low-level westward (upper-level eastward) momentum forcing, reducing the
westerlies to the west but increasing easterlies to the east. In contrast, in the case where MCSs
propagate upshear, the positive feedback between vertical shear and eddy momentum transfer
tends to strength both westerlies (easterlies) to the west (east) at the surface. Due to the relatively
stronger modulation by vertical shear strength to the west, the resulting surface-level westerly
winds dominate. In these two cases, both zonal velocity and potential temperature fields exhibit a
front-to-rear tilt, due to the built-in transition from congestus to deep to stratiform convection.

4. Parameterization of the Upscale Impact of MCSs in the Idealized GCM

According to results in Section 3, the upscale impact of westward-moving MCSs under the
modulation of deep heating excess reproduces a persistent propagating MJO analog with a two-
scale structure and realistic variability of precipitation and winds. In contrast, upscale impact of
upshear-moving MCSs under the modulation of vertical shear strength reproduces a significant
WWB. In this section, we provide a basic parameterization of the upscale impact of upshear-
moving MCSs modulated by both deep heating excess and vertical shear strength. We test the
improvement of key features of the MJO analog in the idealized GCM having clear deficiencies.
a. A basic parameterization of upscale impact of MCSs combining upshear momentum and deep heating excess in the GCM

In reality, the maximum allowable deep heating $P_0$ (conceptually similar to CAPE) should mainly influence the magnitude of mesoscale heating, while the vertical shear strength influences the vertical tilting angles of MCSs (i.e., relative location among shallow congestus, deep and stratiform convection). According to previous results based on the MESD model (Yang and Majda 2018b), both conditions control the magnitude and sign of the eddy transfer of momentum and temperature. Here we combine these two conditions by summing them linearly with a tuning coefficient $\alpha$, and assume that the MCSs all propagate upshear.

A basic parameterization for upscale impact of MCSs (eddy transfer of momentum and temperature $F^u, F^\theta$) is,

$$ F^u = \kappa^u \left( \alpha \frac{P_0}{\bar{Q}} + (1 - \alpha) \frac{|\Delta U|}{U_{ref}} \right) \text{sign}(\Delta U) \left[ -\frac{3}{2} \cos(z) \right] $$

$$ F^\theta = \kappa^\theta \left( \alpha \frac{P_0}{\bar{Q}} + (1 - \alpha) \frac{|\Delta U|}{U_{ref}} \right) \left[ \frac{3}{2} \sin(z) \right] $$

where $P_0 = \frac{1}{z_{conv}} (a_1 \theta_{eb} + a_2 q - a_0 (\theta_1 + \gamma_2 \theta_2))^+$ is the positive excess of the maximum allowable deep heating and $\bar{Q}$ is its RCE value. $\Delta U$ represents the vertical shear strength. $U_{ref} = 10ms^{-1}$. The magnitude coefficients $\kappa^u > 0, \kappa^\theta < 0$ satisfy the relation $\frac{\kappa^\theta}{\kappa^u} = c$, where $c$ is the absolute propagation speed of the MCSs. The coefficient $\alpha$ controls the relative importance of $P_0$ and vertical shear strength in modulating the strength of the eddy transfer of momentum and temperature.

b. Three-way interaction between MJO analog, parameterized upscale impact of MCSs, and background vertical shear on longer time scales

Here we test effects of the parameterization by adding it into the idealized GCM having clear deficiencies. Four cases with various value of $\alpha$ in Eqs.14-15 are considered. The magnitude
coefficient $\kappa^u$ is fixed at 0.0008 and speed of MCSs $c$ is 0.1 (corresponds to 5 $ms^{-1}$). In order to explore the solutions over longer time scales, we extend the integration period to 7000 days and use the last 3000-day model output for analysis. For better visualization, we perform a low-pass filtering by transforming solutions into power spectra in Fourier space and only keeping small wavenumber and frequency (large wavelength and period). Only precipitation anomalies at the length scale longer than 10,000 km and time scale longer than 30 days are retained.

Fig.11 shows the Hovmoller diagrams for precipitation in the cases with various value of $\alpha$. A large value of $\alpha$ corresponds to the case with stronger modulation effects by deep heating excess $P_0$, while a smaller value of $\alpha$ corresponds to the case with stronger modulation effects by vertical shear strength. One particular interesting feature is the direction switching of the MJO analog in the case with $\alpha = 0.8$ (panel c) and $\alpha = 1.0$ (panel d). In panel c, the MJO analog persistently propagates eastward between day 4000 and day 4500, switches to westward propagation between day 4500 and day 4800, then switches back to eastward propagation between day 5000 and day 5300, and so forth. The period between two eastward (or westward) propagation phases is about 800 days, much longer than intraseasonal time scale. Such a QBO-like behavior in the presence of CMT are quite similar to Majda and Stechmann (2009) which also shows periodic direction switching of unstable CCEWs and background mean flow. Compared with panel c, the solution in panel d differs in the duration of persistent propagation of the MJO analog/reversed MJO analog in each phase, exhibiting more chaotic features. For example, the persistently eastward-propagating MJO analog lasts 1200 days between day 5500 and day 6700, while that between day 4500 and day 5000 only last 500 days. Unlike panels c and d, the solutions in panels a and b show little QBO-like behavior. Such a clear difference among the cases with large/small value of $\alpha$ indicates the crucial modulation effects of deep heating excess on the eddy terms from upshear-moving MCSs.
The case with $\alpha = 0.8$ in Fig.12a shows a periodic direction-switch between eastward-propagating MJO analog and westward-propagating reversed MJO analog. Panel b shows the domain-mean zonal winds in the 1st-baroclinic mode, which also exhibits a periodic direction-switch between easterlies and westerlies. Such a QBO-like behavior in domain-mean flow also occurs in the CRM studies by (Held et al. 1993). Specifically, during the phase with eastward-moving (westward-moving) MJO analog, the domain-mean zonal winds gradually increase from low-level easterlies (westerlies) to low-level westerlies (easterlies), reaching its maximum magnitude as the MJO analog switches direction. The persistently eastward (westward) propagation phase is highly correlated with the increasing (decreasing) background zonal winds. According to the governing equations for $u_1$ in Table.1, domain-mean zonal winds vanish in the cases without eddy momentum transfer. Thus, the accumulating contribution by eddy momentum transfer modulated by deep heating excess associated with the MJO analog induces these nonzero domain-mean background flow. Fig.12c shows the time series of domain-mean thermodynamical fields, including 1st-baroclinic potential temperature, boundary-layer equivalent potential temperature, and moisture. The domain-mean 1st-baroclinic potential temperature decreases at each phase when MJO analog persistently propagates westward/eastward. Such cooling effects can be explained by eddy heat transfer from MCSs that accumulate in space and time as the MJO analog persistently propagates across the domain.

Fig.13a shows the zonal/vertical cross-sections of zonal velocity and zonal profiles of deep heating excess and vertical shear strength in the composite eastward-moving planetary-scale envelopes. A significant WWB is reproduced, resembling the realistic MJO analog in Fig.1. A crucial feature is the displacement of the peak of deep heating excess to the west of the dashed line, which is consistent with the observation that convective center of the MJO typically sits over the WWB in easterly vertical shear. Such westward displacement of the deep heating excess preferably mod-
ulates eddy momentum transfer in the trailing edge, resulting in a stronger low-level eastward momentum forcing in the trailing edge than in the leading edge. The relatively weak maximum zonal velocity compared to the realistic MJO analog in Fig.1 is due to the intermittent property of the solutions shown in the Hovmoller diagram in Fig.13b.

We identify the following three-way interaction between MJO analog, parameterized upscale impact of MCSs, and background vertical shear:

1. Eastward-moving MJO analog modulates eddy momentum transfer mainly through deep heating excess.

2. Due to the westward displacement of the deep heating excess, the resulting eddy momentum transfer with low-level eastward momentum forcing accumulates in space and time and switches the low-level background flow from easterlies to westerlies.

3. Background vertical shear with low-level westerlies favors the westward-moving reversed MJO analog. The underlying mechanism is related to eastward moisture advection, resulting in eastward-moving synoptic-scale disturbances and a westward-moving planetary-scale envelope.

4. Mechanisms similar to 1-3 are repeated, but in opposite directions.

Fig.13c-d shows the log-scale power spectra of precipitation and zonal velocity, which is akin to the realistic MJO analog in Fig.1. The power spectra of both fields show a clear peak for eastward-moving planetary-scale envelope at wavenumber 2 and period of 50 days, with a band of extra power extending to higher wavenumber and frequency. For westward-moving components, the power spectra of precipitation shows a peak at wavenumber 5-8 and period of 25-40 days. Extra bands of power spectra occur at higher wavenumber and frequency, while that of zonal velocity has a more dominant peak at smaller wavenumber.

It is interesting to question why the scenario with dominate modulation effects by vertical shear does not exhibit such a QBO-like behavior, considering that the easterly vertical shear in the
trailing edge is stronger than in the leading edge. Although the magnitude of westerly vertical shear in the leading edge is weaker, it covers much broader area. After the eddy momentum transfer in both leading and trailing edges accumulate in space, the resulting background zonal winds are comparable to each other with no persistent direction preference.

5. Concluding Discussion

A simple multicloud model for MJO analog and intraseasonal variability above the equator is studied. With reduced congestus and stratiform heating, the resulting solutions from this simple model are used as an idealized GCM having clear deficiencies. By adding eddy transfer of momentum and temperature predicted by the MESD model, we assess the upscale impact of MCSs on three key features of the MJO analog: persistent propagation of a two-scale structure, realistic planetary/intraseasonal variability in precipitation and winds, and a significant WWB. We then introduce a basic parameterization of upscale impact of upshear-moving MCSs modulated by the effects of deep heating excess and vertical shear strength and test its effects in the idealized deficient GCM.

Table 5 summarizes results reported in this paper regarding the above three key features of the MJO analog in the idealized deficient GCM. Compared to the realistic MJO analog, the idealized deficient GCM fails to reproduce these three features, thereby mimicking the significant bias of the simulated MJO in present-day GCMs. According to Khouider et al. (2012), MCSs and squall lines are prominent in the convectively active regions of the MJO envelope, indicating the modulation of the MCSs by the MJO convective center. The eddy transfer of momentum and temperature from westward-moving MCSs at a slow speed (5 m s$^{-1}$) improves the two-scale structure of the eastward-moving MJO analog and space-time variability of precipitation and winds, but fails to strengthen WWB. This is consistent with the theoretical prediction by the MESD model (Yang...
and Majda 2018b); i.e., westward-moving MCSs embedded in the large-scale convective envelope provide favorable conditions for convection to the east, that promotes the eastward-moving convective envelope. On the other hand, vertical shear plays a crucial role in organized tropical convection (Moncrieff 1992), including the influence on its front-to-rear tilt structure and propagation directions (Moncrieff and Liu 1999; Stechmann and Majda 2009). In particular, eddy transfer of momentum and temperature from upshear-moving MCSs induces a significant WWB in the middle and west of the MJO analog. This is due to the two-way feedback between environmental easterly vertical shear and the embedded eddy momentum transfer with low-level eastward momentum forcing. The eddy transfer of momentum and temperature modulated by the effects of vertical shear strength alone fails to reproduce the two-scale structure of the MJO analog and a realistic space-time variability of precipitation and winds.

In order to incorporate those improvements in global models, we provide a basic parameterization of upscale impact of upshear-moving MCSs that linearly combines the modulation effects of deep heating excess and vertical shear strength. This basic parameterization shares goals similar to the MCSP introduced by Moncrieff et al. (2017); notably representing the upscale effects of organized tropical convection that are missing from contemporary parameterizations in GCMs. The main purpose of the Moncrieff et al. (2017) prototype version of MCSP was to demonstrate the upscale effects of top-heavy convective heating and momentum transport in the simplest possible manner, in order to provide proof-of-concept. This was achieved by focusing on eastward propagation and a full GCM. The results of this present paper will be valuable for the future development of MCSP, because the heating and CMT (i.e., upscale impact of MCSs) have been quantified in simplest ways. However, this basic parameterization differs from the MCSP in several aspects that significantly improve the feasibility and reliability of the parameterization. First, it considers both deep heating excess (a similar concept as CAPE) and vertical shear strength in modulating the
upscale impact of MCSs, while the parameterized CMT in MCSP is simply set as constant magni-

tude over convective regions. Secondly, it assumes eddy transfer of momentum and temperature
from MCSs that propagate upshear (opposite to vertical shear direction), allowing vertical shear to
determine the propagation direction of MCSs and the sign of eddy momentum transfer. Thirdly,
it highlights the crucial contribution of eddy transfer of temperature as predicted theoretically by
the MESD model.

The implementation of this basic parameterization of upscale impact of MCSs in the idealized
deficient GCM shows significant improvement in capturing key features of the MJO. A further ex-
amination of a longer-period simulation reveals a three-way interaction between the MJO analog,
the parameterization of upscale impact of MCSs, and the background mean flow. The westward-
displaced deep heating excess in the eastward-moving MJO analog favors eddy momentum trans-
fer with low-level eastward (upper-level westward) momentum forcing. The effects of the eddy
momentum transfer accumulate in space and time and gradually switches the direction of back-
ground mean flow which, in turn, alter the propagation directions of the MJO analog. Under this
three-way interaction mechanism, the background mean flow exhibits a QBO-like behavior, re-
sembling similar phenomenon in CRM simulations (Held et al. 1993). Although in reality the
Coriolis force would break down the zonal symmetry, such a three-way interaction mechanism
may shed light on the interactions between eastward-moving MJO, upscale impact of MCSs and
climatological vertical shear.

The basic parameterization of upscale impact of MCSs can be elaborated in several ways and
tested in a hierarchy of models. Besides the first-baroclinic mode, it is also interesting to investi-
gate the effects of eddy transfer of momentum and temperature due to high baroclinic modes, as
shown by studies based on the MESD model (Yang and Majda 2018b) and reanalysis data (Oh et al.
2015). A different scenario to assess the upscale impact of MCSs on the planetary/intraseasonal
variability includes the Walker circulation over the warm pool. Furthermore, we would like to test effects of this basic parameterization of upscale impact of MCSs in more comprehensive GCMs.

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Table 1. Prognostic governing equations in the 2D multicloud model for the MJO analog and intraseasonal variability above the equator.

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<th>Name</th>
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<td>Momentum, $j$th-baroclinic mode, $j = 1, 2$</td>
<td>$\frac{\partial u_j}{\partial t} = \frac{\partial \theta_1}{\partial x} - \frac{C_{2} \theta_2}{h_b} u_j - \frac{1}{U} u_j F_j$</td>
</tr>
<tr>
<td>Potential temperature, 1st-baroclinic mode</td>
<td>$\frac{\partial \theta_1}{\partial t} = P - Q_{K,1}^0 - \frac{1}{U} \theta_1 + F_{1}^\theta$</td>
</tr>
<tr>
<td>Potential temperature, 2nd-baroclinic mode</td>
<td>$\frac{\partial \theta_2}{\partial t} - \frac{1}{4} \frac{\partial u_1}{\partial x} = -H_s + H_c - Q_{K,2}^0 - \frac{1}{U} \theta_2 + F_{2}^\theta$</td>
</tr>
<tr>
<td>Free tropospheric moisture</td>
<td>$\frac{\partial \theta}{\partial t} + \frac{\partial}{\partial x} \left[ (u_1 + \tilde{\alpha} u_2) q + \tilde{Q} \left( u_1 + \tilde{\lambda} u_2 \right) \right] = -\frac{1}{U} \beta^2 p + \frac{D}{H_T}$</td>
</tr>
<tr>
<td>Boundary-layer equivalent potential temperature</td>
<td>$\frac{\partial \theta_{eb}}{\partial t} = \frac{1}{h_b} (E - D)$</td>
</tr>
<tr>
<td>Congestus heating</td>
<td>$\frac{\partial H_c}{\partial t} = \frac{1}{\beta} \left( \alpha_c \frac{\Lambda - \Lambda^<em>}{1 - \Lambda^</em>} \frac{D}{H_T} - H_c \right)$</td>
</tr>
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<td>Stratiform heating</td>
<td>$\frac{\partial H_s}{\partial t} = \frac{1}{\beta} (\alpha_s p - H_s)$</td>
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<table>
<thead>
<tr>
<th>Name</th>
<th>Equation</th>
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<tbody>
<tr>
<td>Mid-tropospheric equivalent potential temperature</td>
<td>$\theta_{em} = q + \frac{2\sqrt{2}}{\pi} (\theta_1 + \alpha z \theta_2)$</td>
</tr>
<tr>
<td>Switch function for mid-tropospheric dryness</td>
<td>$\Lambda = \begin{cases} 1 &amp; \text{if } \bar{\theta}<em>{eb} - \bar{\theta}</em>{em} + \theta_{eb} - \theta_{em} \geq 20 K \ \Lambda^+ &amp; \text{if } \bar{\theta}<em>{eb} - \bar{\theta}</em>{em} + \theta_{eb} - \theta_{em} \leq 10 K \end{cases}$</td>
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<tr>
<td>Deep heating</td>
<td>$P = \frac{1 - \Lambda}{1 - \Lambda^+} \rho_0$</td>
</tr>
<tr>
<td>Downdrafts</td>
<td>$= \frac{1 - \Lambda}{1 - \Lambda^+} \left[ \bar{Q} + \frac{1}{\tau_{conv}} (a_1 \theta_{eb} + a_2 q - a_0 (\theta_1 + \gamma_2 \theta_2)) \right]^+$</td>
</tr>
<tr>
<td>$D = \Lambda D_0$</td>
<td></td>
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<tr>
<td>Surface evaporation flux</td>
<td>$\vec{E}<em>h = \frac{1}{2} \left( \bar{\theta}</em>{eb} - \bar{\theta}<em>{em} - \theta</em>{eb} \right)$</td>
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</table>
TABLE 3. Parameters and constants in the idealized GCM with clear deficiencies. The different value of parameters and constants used for realistic MJO analog above the equator is shown in the bracket. All the remaining ones are kept the same as Majda et al. (2007).

<table>
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<tr>
<th>Name</th>
<th>Symbol</th>
<th>Value</th>
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<tbody>
<tr>
<td>First baroclinic radiative cooling rate</td>
<td>$Q_{R,1}^0$</td>
<td>1 K/day</td>
</tr>
<tr>
<td>Stratiform adjustment coefficient</td>
<td>$\alpha_s$</td>
<td>0.125 (0.25)</td>
</tr>
<tr>
<td>Congestus adjustment coefficient</td>
<td>$\alpha_c$</td>
<td>0.25 (0.5)</td>
</tr>
<tr>
<td>Height of troposphere</td>
<td>$H_T$</td>
<td>15.7 km</td>
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<tr>
<td>Height of the boundary layer</td>
<td>$h_b$</td>
<td>500 m</td>
</tr>
<tr>
<td>Relative contribution of stratiform and congestus to downdrafts</td>
<td>$\mu_2$</td>
<td>0.5</td>
</tr>
<tr>
<td>Convective time scale</td>
<td>$\tau_{conv}$</td>
<td>12 hrs</td>
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<tr>
<td>Momentum drag time scale due to turbulent fluctuations</td>
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<td>28.9 days</td>
</tr>
<tr>
<td>Rayleigh-wind relaxation time scale</td>
<td>$\tau_R$</td>
<td>150 days</td>
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<td>$\tau_D$</td>
<td>100 days</td>
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<tr>
<td>Stratiform adjustment time scale</td>
<td>$\tau_s$</td>
<td>7 days</td>
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<td>Congestus adjustment time scale</td>
<td>$\tau_c$</td>
<td>7 days</td>
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<td>Inverse convective buoyancy time scale of deep clouds</td>
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<td>12</td>
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<td>$a_1$</td>
<td>0.1</td>
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<tr>
<td>Relative contribution fraction of $q$ to deep convection</td>
<td>$a_2$</td>
<td>0.9</td>
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</tr>
<tr>
<td>Coefficient of $v_2$ in linear moisture convergence</td>
<td>$\check{\lambda}$</td>
<td>0.6</td>
</tr>
<tr>
<td>Background moisture stratification</td>
<td>$\check{\rho}$</td>
<td>1.03 (1.0)</td>
</tr>
<tr>
<td>Lower threshold of the switch function $\Lambda$</td>
<td>$\Lambda^*$</td>
<td>0.2</td>
</tr>
</tbody>
</table>
**Table 4.** Value of thermodynamic variable at RCE state. The remaining variables not mentioned here are all zero at RCE state. The different value of parameters and constants used for realistic MJO analog above the equator is shown in the bracket.

<table>
<thead>
<tr>
<th>Name</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Discrepancy between boundary layer and middle troposphere $\theta_e$</td>
<td>$\bar{\theta}<em>{eb} - \bar{\theta}</em>{em}$</td>
<td>12 K</td>
</tr>
<tr>
<td>Discrepancy between boundary layer $\theta_e$ and its saturated value</td>
<td>$\theta_{eb}^* - \bar{\theta}_{eb}$</td>
<td>10 K</td>
</tr>
<tr>
<td>Moisture switch at RCE</td>
<td>$\Lambda$</td>
<td>0.36</td>
</tr>
<tr>
<td>Bulk convective heating at RCE</td>
<td>$\bar{Q}$</td>
<td>1.25 K day$^{-1}$</td>
</tr>
<tr>
<td>Congestus heating at RCE</td>
<td>$\bar{H}_c$</td>
<td>0.045 K day$^{-1}$ (0.09 K day$^{-1}$)</td>
</tr>
<tr>
<td>Deep heating at RCE</td>
<td>$\bar{H}_d$</td>
<td>1 K day$^{-1}$</td>
</tr>
<tr>
<td>Stratiform heating at RCE</td>
<td>$\bar{H}_s$</td>
<td>0.125 K day$^{-1}$ (0.25 K day$^{-1}$)</td>
</tr>
<tr>
<td>Second baroclinic radiative cooling rate</td>
<td>$\bar{Q}_{R,2}$</td>
<td>-0.08 K day$^{-1}$ (-0.16 K day$^{-1}$)</td>
</tr>
<tr>
<td>Downdraft mass flux reference scale</td>
<td>$m_0$</td>
<td>0.0364 ms$^{-1}$ (0.035 ms$^{-1}$)</td>
</tr>
<tr>
<td>Evaporation time scale</td>
<td>$\tau_e$</td>
<td>8.49 hrs</td>
</tr>
</tbody>
</table>
TABLE 5. Summary of all experiments under the different model setup and their results in capturing key features of the MJO. In the upscale impact of MCSs column, no means no eddy is added, westward/eastward means the propagation direction of MCSs, slow/fast corresponds to $5/20 \, ms^{-1}$, upshear/downshear means the propagation direction of MCSs is opposite/along vertical shear direction. The ”modulation” column shows the modulation effects of deep heating excess $P_0$ and vertical shear strength. The ”key feature” column includes (1) two-scale structure of the MJO analog, (2) power spectra of precipitation and winds with planetary/intraseasonal peaks, (3) westerly wind burst.

<table>
<thead>
<tr>
<th>#</th>
<th>Model Setup</th>
<th>Key Feature</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>upscale impact of MCSs</td>
<td>modulation</td>
<td>two-scale</td>
</tr>
<tr>
<td>Realistic MJO analog</td>
<td>no</td>
<td>N/A</td>
<td><strong>good</strong></td>
</tr>
<tr>
<td>Idealized GCM with clear deficiencies</td>
<td>no</td>
<td>N/A</td>
<td>bad</td>
</tr>
<tr>
<td>westward, slow</td>
<td>$P_0$</td>
<td><strong>good</strong></td>
<td><strong>good</strong></td>
</tr>
<tr>
<td>westward, fast</td>
<td>$P_0$</td>
<td>bad</td>
<td><strong>good</strong></td>
</tr>
<tr>
<td>westward shear</td>
<td>shear</td>
<td><strong>good</strong></td>
<td>bad</td>
</tr>
<tr>
<td>upshear shear</td>
<td>bad</td>
<td>bad</td>
<td><strong>good</strong></td>
</tr>
<tr>
<td>downshear shear</td>
<td>bad</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>upshear $P_0$ &amp; shear</td>
<td><strong>good</strong></td>
<td><strong>good</strong></td>
<td>good</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

Fig. 1. Realistic MJO analog above the equator. Hovmoller diagram for (a) precipitation and log-scale power spectra for (b) precipitation and (c) surface-level zonal velocity, (d-e) vertical cross-sections of composite planetary-scale envelope in the moving reference frames (6.1 ms\(^{-1}\)), based on model output between day 3000 and day 4000. Panels (d) shows PH, \(H_e\) with the left y-axis and \(q, \theta_{eb}\) with the right y-axis and panel (e) shows zonal velocity (color) and potential temperature (contour, solid lines for positive value, dashed lines for negative, contour interval 0.05K). The pink curve shows the zonal profile of precipitation anomalies with the right axis. Domain-mean potential temperature is removed. The units of precipitation and zonal velocity are \(K_{day}^{-1}, ms^{-1}\), respectively.

Fig. 2. An idealized GCM with clear deficiencies. Hovmoller diagram for (a) precipitation and log-scale power spectra for (b) precipitation and (c) surface-level zonal velocity based on model output between day 3000 and day 4000. The units of precipitation and zonal velocity are \(K_{day}^{-1}, ms^{-1}\), respectively.

Fig. 3. Vertical profiles of (a) zonal/vertical velocity (arrows), (b) potential temperature anomalies (contours; interval: 0.06 K), (c) eddy momentum transfer and (d) eddy heat transfer in an eastward-moving mesoscale system. The colors in panels (a,b) show mesoscale heating. The maximum magnitudes of zonal and vertical velocities are 3.73\(ms^{-1}\) and 0.47\(ms^{-1}\), respectively. Panels (c,d) also show truncated eddy transfer of momentum and temperature with only the 1st-baroclinic mode. One dimensionless unit of eddy momentum transfer and eddy heat transfer is 15 \(ms^{-1}day^{-1}\) and 4.5 \(K_{day}^{-1}\), respectively.

Fig. 4. Hovmoller diagrams for precipitation between day 3800 and day 4000 in the cases with/without eddy terms from westward-moving MCSs modulated by \(P_0\). Panel (a) shows the case without eddy terms. Panels (b-c) show the case with eddy terms from (b) slowly propagating MCSs (5 \(ms^{-1}\)) and (c) fast propagating MCSs (20 \(ms^{-1}\)). The unit is \(K_{day}^{-1}\).

Fig. 5. Log-scale power spectra of precipitation (left column) and zonal velocity (right column) in the wavenumber-frequency diagrams, based on the model output between day 3000 and day 4000 in the cases with/without eddy terms from westward-moving MCSs modulated by \(P_0\). The rows from top to bottom correspond to the case with (a,b) no eddy, (c,d) eddy terms from slowly propagating MCSs (5 \(ms^{-1}\)), (e,f) eddy terms from fast propagating MCSs (20 \(ms^{-1}\)). Each column share the same color as placed in the bottom.

Fig. 6. Vertical cross-sections of composite planetary-scale envelope in the moving reference frames (s is the propagating speed), based on model output between day 3000 and day 4000 in the cases with/without eddy terms from westward-moving MCSs modulated by \(P_0\). Panels (a-b) show the cases with eddy terms from (a) slowly propagating MCSs (5 \(ms^{-1}\)) and s=3.05 \(ms^{-1}\), and (b) fast propagating MCSs (20 \(ms^{-1}\)) and s=3.35 \(ms^{-1}\). Zonal velocity is shown by color and potential temperature is shown by contours (solid lines for positive value, dashed lines for negative, contour interval 0.005K). The pink curve shows the zonal profile of precipitation anomalies with the right axis. Domain-mean potential temperature is removed. The units of zonal velocity and potential temperature are \(ms^{-1}, K\), respectively.

Fig. 7. A conceptual diagram for the definition of (a) vertical shear strength and (b) upshear/downshear propagation. In panel (a), the red (blue) bars indicate the maximum (minimum) magnitude of zonal winds in the upper and lower tropospheres. The strength of vertical shear is defined as the stronger magnitude between westerly and easterly vertical shear, \(\max\left\{\left|U_{max}^u - U_{min}^u\right|, \left|U_{min}^u - U_{max}^l\right|\right\}\), and its direction is determined correspondingly.
In panel (b), upshear (downshear) is defined as propagation along the opposite (same) direction of vertical shear.

Fig. 8. Similar to Fig.4 but the cases with/without eddy terms modulated by vertical shear strength. Panel (a) shows the case without eddy terms. The remaining panels shows the case with eddy terms from MCSs propagating (b) westward, (c) upshear, (d) downshear.

Fig. 9. Similar to Fig.5 but for the cases with/without eddy terms modulated by vertical shear strength. The rows from top to bottom correspond to the case with (a,b) no eddy, (c,d) eddy terms from westward-moving MCSs, (e,f) eddy terms from upshear-moving MCSs.

Fig. 10. Similar to Fig.6 but the cases with/without eddy terms modulated by vertical shear strength. Panel (a-b) show the cases with eddy terms from (a) westward-moving MCSs and s=3.075 $ms^{-1}$, and (b) upshear-moving MCSs and s=3.5 $ms^{-1}$.

Fig. 11. Hovmoller diagrams for planetary/intraseasonal anomalies (deviation from RCE value) of precipitation between day 4000 and day 7000. These panels correspond to the cases with $\alpha$ equal to (a) 0.0, (b) 0.4, (c) 0.8, (d) 1.0. The planetary/intraseasonal anomalies are obtained by using a low-pass filter and only those on length scale larger than 10000 km and time scale longer than 30 days are retained. The unit is $Kday^{-1}$.

Fig. 12. Time series of precipitation, zonal velocity and thermodynamical fields between day 4000 and day 7000. Panel (a) shows the Hovmoller diagram for planetary/intraseasonal anomalies of precipitation (the same as Fig.11c), while panel (b) and (c) show domain-mean zonal velocity and thermodynamic fields ($\theta_1, \theta_{eb}, q$) during the same period, respectively. Only anomalies of these thermodynamic fields ($\theta_1, \theta_{eb}, q$) on the time scale longer than 50 days are regained by using the low-pass filter. The units of precipitation and zonal velocity are $Kday^{-1}, ms^{-1}$, respectively.

Fig. 13. An idealized GCM with clear deficiencies and extra parameterization for upscale impact of MCSs with $\alpha = 0.8$. Panel (a) shows vertical cross-sections of composite planetary-scale envelope in the moving reference frames ($3.65 ms^{-1}$) based on model output between day 5750 and day 6000. Zonal velocity is shown by color. Dimensionless value of deep heating excess $\frac{\mathcal{H}}{Q}$ is shown by pink curve, while that of vertical shear strength $\frac{|\Delta U|}{U_{ref}}$ is shown by black curve. The dashed line indicate the longitude with easterly (westerly) vertical shear to its west (east). Panel b shows the Hovmoller diagram for precipitation during this 250-day period. Panels (b-c) show log-scale power spectra for (b) precipitation and (c) surface-level zonal velocity. The units of precipitation and zonal velocity are $Kday^{-1}, ms^{-1}$, respectively.
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