Role of Diurnal Cycle in the Maritime Continent Barrier Effect on MJO Propagation in an AGCM

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ABSTRACT: The barrier effect of the Maritime Continent (MC) in stalling or modifying the propagation characteristics of the MJO is widely accepted. The strong diurnal cycle of convection over the MC is believed to play a dominant role in this regard. This hypothesis is studied here, with the help of a coarse-resolution atmospheric general circulation model (AGCM). The dry dynamical core of the AGCM is coupled to the multcloud parameterization piggybacked with a dynamical bulk boundary layer model. A set of sensitivity experiments is carried out by systematically varying the strength of the MC diurnal flux to assess the impact of the diurnal convective variability on the MJO propagation. The effects of deterministic and stochastic diurnal forcings on MJO characteristics are compared. It is found that the precipitation and zonal wind variance, on the intraseasonal time scales, over the western Pacific region decreases with the increase in diurnal forcing, indicating the blocking of MC precipitation. An increase in precipitation variance over the MC associated with the weakening of precipitation variance over the west Pacific is evident in all experiments. The striking difference between deterministic and stochastic diurnal forcing experiments is that the strength needed for the deterministic case to achieve the same degree of blocking is almost double that of stochastic case. The stochastic diurnal flux over the MC seems to be more detrimental in blocking the MJO propagation. This hints at the notion that the models with inadequate representation of organized convection tend to suffer from the MC-barrier effect.

KEYWORDS: Maritime Continent; Madden-Julian oscillation; General circulation models; Diurnal effects; Intraseasonal variability

1. Introduction

The Maritime Continent (MC) is a unique geographical region, where large-scale interaction between land and ocean occurs due to the existence of thousands of small islands, mountainous terrain, and shallow seas. This region is also known as the Indo-Pacific warm pool, where sea surface temperatures (SSTs) higher than 28°C persist through the year. The presence of moisture and the unstable environment favors the development of a deep convective activity that is deemed to have a significant impact on the propagation of Madden-Julian oscillation (MJO) disturbances (Salby and Hendon 1994; Zhang and Hendon 1997; Kerns and Chen 2016) from the Indian Ocean to the western Pacific. It has been widely hypothesized that, in this fashion, the MC acts as a barrier for the eastward propagation of MJO disturbances (Rui and Wang 1990; Salby and Hendon 1994; Zhang and Ling 2017). Interactions across scales between the diurnal cycle of convection over the MC and the MJO dynamics has been considered as the possible culprit by many recent studies (Neale and Slingo 2003; Peatman et al. 2014; Birch et al. 2016; Hagos et al. 2016; Majda and Yang 2016; Sakaeda et al. 2017; Lu et al. 2019).

The behavior of MJO related convection undergoes considerable changes as it moves from the Indian Ocean (IO) to the Pacific Ocean through the MC. Observational studies suggest that the propagation speed of the MJO weakens over the MC and sometimes to the point of stalling it over the islands (Hendon and Salby 1994; Peatman et al. 2014; Kerns and Chen 2016). A detailed insight into the complex processes that cause the blocking of MJO as it interacts with the MC is vital for simulating the MJO and convection activity over the MC in general circulation models (GCM). Studies have shown that models have poor skill in simulating the barrier effect of MC (e.g., Chen et al. 2020). Vitart and Molteni (2010) shows that while 30% of MJO events fail to propagate through MC in a reanalysis product, 50% of events fails to pass MC in the European Centre for Medium-Range Weather Forecasts (ECMWF) forecast system. In general, GCMs struggle to reproduce the complex precipitation characteristics over the MC (Seo et al. 2009; Weaver et al. 2011; Neena et al. 2014; Peatman et al. 2016). Apart from causing variations in the structure of the MJO disturbances (Kiladis et al. 2005), the interaction with MC also influences the local climate by modulating the local precipitation and cloudiness, especially over the islands (Rauniyar and Walsh 2011; Oh et al. 2012; Moron et al. 2015; Sakaeda et al. 2017).

Studies suggest several possible reasons for the stalling of MJO disturbances over the MC. The heat fluxes over the ocean and atmosphere play a fundamental role in driving the MJO (e.g., Sobel et al. 2008). The reduced latent heat flux over the MC due to the presence of many islands reduces moisture

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source for MJO and can weaken the MJO (Maloney and Sobel 2004; Sobel et al. 2008). Another source of instability is cloud–radiation feedback (nonlinear interaction among radiation, convection, and surface fluxes) that can influence the smooth propagation of MJOs by modulating longwave cooling by cloud and moisture anomalies (Hu and Randall 1994; Sobel and Maloney 2013). The topographic influence of the MC can distort the moisture convergence of low-level circulation and can hinder the propagation of the MJO (Hsu and Lee 2005; Wu and Hsu 2009). The land convection over MC plays a key role in shaping the moisture distribution around the MC through its local and nonlocal effects (Ling et al. 2019; Ahn et al. 2020). A strong wet bias in the vicinity of the MC or a cold SST bias in the equatorial Pacific can cause unfavorable conditions for MJO propagation (Kim et al. 2016; DeMott et al. 2018).

The strong diurnal variability over the MC is another major factor that influences the propagation of MJO disturbances (Neale and Slingo 2003; Tung et al. 2014; Hagos et al. 2016; Majda and Yang 2016; Ling et al. 2019). Majda and Yang (2016) show that planetary-scale diurnal variability of diabatic heating can have an upscale influence on intraseasonal variability through eddy flux divergence of potential temperature. Specifically, the temperature anomaly induced by the intraseasonal envelope of the diurnal cycle can offset the effect of convective momentum transfer on the MJO, yielding stalled or suppressed propagation of the MJO across the MC. Despite the unveiling of the various plausible factors responsible for MC barrier, an explanation on why some MJO events stall over the MC while others cross the MC is lacking. More clarity in this regard is needed for realistic simulation of MJO in numerical models.

As mentioned above, some MJOs stall or weaken over the MC whereas others remain intact (e.g., Kerns and Chen 2016). The MC diurnal cycle has been considered as the possible reason for the barrier effect (e.g., Hagos et al. 2016). In this study, we examine the impact of the diurnal variability of MC on the MJO propagation using a coarse-resolution atmospheric general circulation model (AGCM). This study specifically focuses on the role of the strength of the diurnal cycle (by varying the sensitivity of surface flux) of convection in stalling the MJO over the MC. This is a unique component of the present study in comparison to others (e.g., Hagos et al. 2016; Ruppert and Johnson 2016) who often remove the diurnal cycle of solar radiation to modulate the diurnal cycle of rainfall. In some sense, the present study is the converse of many previous studies that showed that different MJO phases are associated with different strength regimes of the diurnal convection over the MC. We expect to gain insight into how the natural variability of the diurnal cycle over MJO in terms of its strength and timing can influence the MJO propagation with the help of a coarse-resolution GCM.

Specifically, we impose a diurnally varying surface flux over the MC islands whose maximum strength is systematically changed from a very weak diurnal forcing to a strong diurnal forcing. The GCM consists of the High-Order Methods Modeling Environment (HOMME) atmospheric dynamical core coupled to a simple multicloud convective parameterization (Khouider et al. 2011) and a bulk planetary boundary

layer model (Stevens 2006; Waite and Khouider 2009). The coupling of the multicloud convective parameterization and the bulk boundary layer model has proven to realistically capture the diurnal cycle of convection both over land and over the ocean, including the characteristic early morning peak over the ocean and a late afternoon peak over land. The MC diurnal convection variability depends on several factors, such as its modulation of convectively coupled waves, mesoscale systems propagating between land and ocean, sea breeze, and regional winds, to name a few (e.g., Yang and Slingo 2001; Sakaeda et al. 2020). Thus, it is reasonable to assume that the effect of the diurnal cycle on MJO is stochastic. Here, we compare the impact of both deterministic and stochastic diurnal forcings to assess the effect of stochasticity of MC diurnal convection on MJO propagation. Note that our intention is not to mimic the stochastic variability of this phenomenon as in nature. Instead, we use a Poisson process to simulate this forcing stochastically, which is more in line with the pseudoregular and unorganized (grid-scale) character of the diurnal convection as represented by state-of-the-art climate models (Klingaman et al. 2017; Martin et al. 2017; Peters et al. 2017). This study also provides some indications of the cause behind the association between the MJO propagation and the strength of the diurnal convection over the MC islands.

The paper is organized as follows. In section 2, we describe the data, the numerical model, and the simulation setup. Section 3 presents an overview of the impact of enhancing the variability of the MC diurnal convection on the extent of MJO-like variability in the western Pacific, both in deterministic and stochastic diurnal forcing experiments. A summary and a few concluding remarks are given in section 4.

2. Data and computational setup

a. Model description

The computational setup is based on a modified version of the HOMME (Dennis et al. 2005; Nair and Tufo 2007) model dynamical core, using the multicloud model parameterization of convection (Khouider et al. 2011). The main modifications made here include a new treatment of the boundary layer dynamics within the implementation of the multicloud model (MCM) and the use of realistic surface fluxes and background climatology, drawn from reanalysis data. For the boundary layer dynamics, we replaced the original simple thermodynamic bulk boundary layer, reduced to a single equation for the boundary layer potential temperature (θ_{eb}), by a comprehensive bulk boundary layer model (Stevens 2006). This setup was successfully coupled to the multicloud parameterization and used previously in a simplified setting to study the influence of boundary layer dynamics on convectively coupled waves and the diurnal cycle (Waite and Khouider 2009; Frenkel et al. 2011a,b). The HOMME model solves the hydrostatic primitive equations on a cubed sphere using a spectral element method (Dennis et al. 2005; Nair and Tufo 2007) and is run here at coarse resolution with a mesh size of roughly 160 km in the horizontal, 26 vertical levels uniformly distributed in σ coordinates, and a 0.5 min time step.

The multicloud model assumes three heating profiles associated with the cloud types, congestus, deep, and stratiform, that characterize tropical convective systems. These heating profiles force the first and second baroclinic modes of the vertical structure directly (Kasahara and Puri 1981). Midlevel moisture is used as a proxy to switch between congestus and deep convection regimes. When the midtroposphere is dry, congestus clouds are favored, and when it is moist, deep convection is preferred. Stratiform heating trails deep convection by a prescribed time lag of 3 h (Khouider and Majda 2006). Congestus clouds serve to moisten the midtroposphere via the induced second baroclinic low-level convergence and precondition the environment for deep convection [see Khouider et al. (2011) for details].

The bulk PBL model is coupled to the MCM convective parameterization following Waite and Khouider (2009) and Frenkel et al. (2011a). The PBL dynamics are represented by equations for the horizontal velocity, potential temperature, and specific humidity averaged over the depth of the PBL. Further details on the MCM and the bulk dynamical boundary layer models can be found in the above mentioned references. For the sake of completeness, the key dynamical equations and variables are reported in Table 1. A list of tunable parameters used by the two models is provided in Table 2.

To take into account the feedback of the PBL dynamics onto the resolved model dynamics, a boundary layer damping term is added to the global model by enforcing the leapfrog update

$$\mathbf{u}^{n+1} = \mathbf{u}^{n-1} + 2\Delta t \left(\frac{E_u^n}{h_b} \Delta_t \mathbf{u}^n \psi(z) \right) \quad (1)$$

at each time step, where n is the time stepping index and Δt is the GCM time step. Here $\psi(z)$ is a weight function, which decays exponentially with the height z above the boundary layer:

$$\psi(z) = \begin{cases} 1 & \text{for } 0 \leq z \leq h_b, \\ e^{-(z-h_b)/h_b} & \text{for } z > h_b, \end{cases}$$

where h_b is the boundary layer height.

In addition to the boundary layer damping, the surface boundary condition for the free-tropospheric vertical velocity is set to $w = -h_b (\nabla \cdot \mathbf{u}_b)$, at the first model level (Waite and Khouider 2009). Note that this is implemented in terms of pressure vertical velocity: $\omega/p = -wg/RT$ where p and T are the first level model pressure and temperature, respectively. The correction to the first model level pressure, due to the presence of the boundary layer represented by the term $-(gh_b/\theta_0)\theta_b$ in the PBL pressure equation (cf. Table 1), is negligible as demonstrated numerically (results not shown).

In the total heating equation (cf. Table 1), ψ_1 , ψ_2^c , and ψ_2^s are the heating vertical profiles associated with deep, congestus, and stratiform clouds, respectively, and H_d , H_c , and H_s are the corresponding heating rates. The functions ψ_1 , ψ_2^c , ψ_2^s are precisely the first and second baroclinic vertical eigen modes of Kasahara and Puri (1981), as indicated by the subscripts 1 and 2, truncated to zero above the 200 hPa level to avoid nonphysical heating or cooling above that level (Khouider et al. 2011). The term $\{\alpha_E(E/H_T)\Delta_t\theta + \beta_D[M_d/H_T + (\nabla \cdot \bar{\mathbf{u}})]\Delta_m\theta\}\psi_1$,

in the free-tropospheric moisture equations, represents the effects of heating and cooling due to shallow cumulus entrainment and detrainment and vertical turbulent mixing where α_E and β_D are tunable parameters while H_T is the tropospheric height (see Table 2).

The factor M_{ask} is a binary function taking the value one inside the equatorial strip between 50°N and 50°S and zero elsewhere, smoothed using the hyperbolic tangent and $\kappa \approx 2/3$ is the ratio of the gas constant and the specific heat capacity at a constant pressure of dry air (Khouider et al. 2011). The last term in convective heating equation accounts for the relaxation of the model temperature to fixed background temperature T_B set by the mean climatology (see below), with a relaxation time scale fixed to $\tau_r = M_{\text{ask}}\tau + (1 - M_{\text{ask}})\tau_f$ so that a long relaxation $\tau_r = \tau = 50$ days is achieved within the tropical strip while a faster relaxation of $\tau_r = \tau_f = 1$ day is used outside, to nudge the model thermodynamics to the background climatology. The same relaxation is applied for the momentum equations as well.

b. Background climatology data as a radiative–convective equilibrium

1) RCE EQUATIONS

A radiative–convective equilibrium (RCE) is a state of rest in the statistical equilibrium sense (long time average) where radiative forcing is balanced by convective heating. The mathematical formulation of the RCE state is easily obtained when considering the governing equations in a steady state of rest.

Eliminating the boundary layer radiative cooling Q_{RB} from the steady state version of the PBL potential and equivalent potential temperatures in Table 1, leads to

$$\left(\frac{1}{\alpha_m - 1} \right) (\Delta_t \bar{\theta}_e - \Delta_t \bar{\theta}_l) - (\Delta_m \bar{\theta}_e - \Delta_m \bar{\theta}) + \frac{h_b}{M_d \tau_e} (\Delta_s \bar{\theta}_e - \Delta_s \bar{\theta}) = 0. \quad (2)$$

Here, the overbars denote longtime averages. We note that some of the quantities, such as the potential and equivalent potential temperature discrepancies can be directly obtained from data. Other parameter values are inferred from the RCE equations, which significantly reduces the number of tuning parameters (Khouider and Majda 2006; Waite and Khouider 2009).

For example, the above equation is used to solve for the RCE downdraft mass flux $\bar{M}_d$. From the free-tropospheric moisture and the convective heating tendency equations (see Table 1), we can obtain an equation for the radiative cooling associated with the first baroclinic mode [Eq. (3)], if we assume $\bar{T} = T_B$ and setting temperature perturbation from this background to zero and the MCM closure equations:

$$Q_{R1}^0 = \frac{\bar{M}_d \Delta_m \bar{\theta}_e}{H_T}, \quad (3)$$

from which we get

$$Q_{R2}^0 = Q_{R1}^0 \left[\alpha_c \frac{\bar{\Lambda} - \Lambda^*}{1 - \bar{\Lambda}} - \alpha_s \right] \quad (4)$$

TABLE 1. Dynamical equations of the multcloud parameterization and bulk dynamical boundary layer. Here, $\Delta_s\phi = \phi_s - \phi_b$, $\Delta_t\phi = \phi_b - \phi_t$, and $\Delta_m\phi = \phi_b - \phi_m$, where the subscripts s, t, b , and m indicate values of the variable ϕ taken at the surface, the top of the PBL, over the bulk of the PBL (vertical average), and the middle of the troposphere, respectively. See text for other details.

Variable	Equation
Bulk PBL equivalent potential temperature	$\frac{\partial\theta_{eb}}{\partial t} + \mathbf{u}_b \cdot \nabla\theta_{eb} = -\frac{E}{h_b}\Delta_t\theta_e - \frac{M_d}{h_b}\Delta_m\theta_e + \frac{1}{\tau_e}\Delta_s\theta_e - Q_{RB}$
Bulk PBL potential temperature	$\frac{\partial\theta_b}{\partial t} + \mathbf{u}_b \cdot \nabla\theta_b = -\frac{E}{h_b}\Delta_t\theta - \frac{M_d}{h_b}\Delta_m\theta + \frac{1}{\tau_e}\Delta_s\theta - Q_{RB}$
PBL entrainment of mass	$E = [M_u - M_d + h_b\nabla \cdot \mathbf{u}_b]^+$
PBL updraft mass flux	$M_u = (1/\alpha_m)D_c$
PBL downdraft mass flux	$M_d = [D_c + h_b\nabla \cdot \mathbf{u}_b]^+$
Bulk PBL horizontal velocity	$\frac{\partial\mathbf{u}_b}{\partial t} + \mathbf{u}_b \cdot \nabla\mathbf{u}_b + f\mathbf{u}_b^\perp = -\nabla p_b - \frac{E_u}{h_b}\Delta_t\mathbf{u} - \frac{C_d U}{h_b}\mathbf{u}_b$
Bulk PBL pressure	$p_b = \frac{p_s + p_{h_b}}{2}$
Top PBL pressure	$p_{h_b} = p_s - \frac{gh_b}{\theta_0}\theta_b$
PBL entrainment of momentum	$E_u = \left[\frac{h_b}{\tau_T} + h_b\nabla \cdot \mathbf{u}_b \right]^+$
Convective heating	$Q_{\text{tot}} \equiv \frac{DT}{Dt} = \frac{1}{1-\kappa} \left[H_d\psi_1 + H_c\psi_2^c - H_s\psi_2^s + \left(\alpha_E \frac{E}{H_T}\Delta_t\theta + \beta_D \left(\frac{M_d}{H_T} + (\nabla \cdot \mathbf{u}) \right) \Delta_m\theta \right) \psi_1 - Q_R \right] \left(\frac{p}{p_0} \right)^k M_{\text{ask}} + \frac{1}{\tau_T} [T_B - T]$
Longwave radiative cooling	$Q_R = Q_{R1}^0\psi_1 + Q_{R2c}^0\psi_2^c - Q_{R2s}^0\psi_2^s$
Free-tropospheric moisture	$\frac{\partial q}{\partial t} = -[\tilde{\mathbf{u}} \cdot \nabla q + \nabla \cdot (\mathbf{u}_1 + \tilde{\alpha}\mathbf{u}_2)q + \nabla \cdot (\mathbf{u}_1\tilde{Q}_1 + \mathbf{u}_2\tilde{Q}_2) - \nabla \cdot \tilde{\mathbf{u}}\tilde{Q}_0] - P + \frac{E}{H_T}\Delta_t q + (1 - \alpha_E)\frac{E}{H_T}\Delta_t\theta + \left(\frac{M_d}{H_T} + \nabla \cdot \mathbf{u} \right) \Delta_m q + (1 - \beta_D) \left(\frac{M_d}{H_T} + \nabla \cdot \mathbf{u} \right) \Delta_m\theta.$
Precipitation	$P = H_d\bar{\psi}_1 + [H_c\bar{\psi}_2^c - H_s\bar{\psi}_2^s]$
Deep convection heating	$H_d = (1 - \Lambda)Q_d$
Congestus heating	$\partial H_c/\partial t = (1/\tau_c)(\alpha_c\Lambda Q_c - H_c)$
Stratiform heating	$\partial H_s/\partial t = (1/\tau_s)(\alpha_s H_d - H_s)$
Deep convective heating potential	$Q_d = \left\{ Q_c^0 + \frac{1}{\tau_{\text{conv}}} [a_1\theta_{eb} + a_2q - a_0(\theta_1 + \gamma_2\theta_2)] \right\}^+$
Congestus heating potential	$Q_c = \left\{ Q_c^0 + \frac{1}{\tau_{\text{conv}}} [\theta_{eb} - a'_0(\theta_1 + \gamma'_2\theta_2)] \right\}^+.$
Moisture switch function	$\Lambda = \begin{cases} 1, & \text{if } \Delta_m\theta_e > \theta^+ \\ A\Delta_m\theta_e + B, & \text{if } \theta_- \leq \Delta_m\theta_e \leq \theta^+ \\ \Lambda^* \equiv 0, & \text{if } \Delta_m\theta_e < \theta^- \end{cases}$
Deep convective downdrafts	$D_0 = D_c(\theta_{eb} - \theta_{em})$ $D_c = m_0 \left[1 + \frac{\mu}{Q} (H_s - H_c) \right]^+$
Midtropospheric equivalent potential temperature	$\theta_{em} = \theta_m + q_{mc}$
Midtropospheric potential temperature	$\theta_m = \psi_1(500 \text{ hPa})(\theta_1 + \alpha_2\theta_2)$

and

$$m_0 = \frac{\bar{M}_d}{1 - \mu \frac{Q_{R2}^0}{Q_{R1}^0}}. \quad (5)$$

Finally, the RCE of the θ_{eb} equation is then used to recover the boundary layer radiative cooling (Q_{RB}). We recall that the tropospheric height is fixed to 15 km while the boundary layer height varies with latitude and

longitude and is set to its climatological value obtained from data.

2) RCE VALUES INFERRED FROM REANALYSIS DATA

Our strategy here is to use the climatological values of temperature and moisture, drawn from reanalysis data, to determine RCE values for the discrepancies Δ_s , Δ_t , and Δ_m of θ and θ_e (see caption of Table 1) and use them to infer the radiative cooling rates, Q_{R1}^0 , Q_{R2}^0 , and Q_{RB} , and m_0 , which in turn are used to force the large-scale model (HOMME-MCM).

TABLE 2. Parameters and variables in HOMME-MCM model.

Parameter	Value	Description
ne	20	Number of elements for HOMME
tstep	30 s	Model time step
nlev	26	Number of vertical levels
mask	0–1	Tropical mask smoothed Heaviside function with slope $k = 10$ and $y_0 = \pm 50$
Q_{RB}	2D varying (Fig. 1i)	Boundary layer radiative cooling
Q_{R1}^0	2D varying (Fig. 1h)	First baroclinic radiative cooling rate
Q_{R2}^0	2D varying (not shown)	Second baroclinic radiative cooling rate
$\tilde{Q}_0$	2D varying (Fig. 1j)	Barotropic projection of the background moisture gradient
$\tilde{Q}_1$	2D varying (Fig. 1k)	First baroclinic projection of the background moisture gradient
$\tilde{Q}_2$	2D varying (Fig. 1l)	Second baroclinic projection of the background moisture gradient
m_0	Determined at RCE	Background downdraft velocity
$\bar{\theta}^s$	2D varying	Background surface θ calculated from ERAI climatology
$\bar{\theta}_e^s$	2D varying	Background surface θ_e calculated from ERAI climatology
$\bar{\theta}^i$	2D varying	Inversion-layer θ calculated from ERAI climatology
$\bar{\theta}_e^i$	2D varying	Inversion-layer θ_e calculated from ERAI climatology
$\bar{\theta}_m$	2D varying	Midtroposphere θ calculated from ERAI climatology at 700 hPa
$\bar{\theta}_e^m$	2D varying	Midtroposphere θ_e calculated from ERAI climatology at 700 hPa
$\bar{\theta}_b$	2D varying	Boundary layer θ calculated from ERAI climatology
$\bar{\theta}_e^b$	2D varying	Boundary layer θ_e calculated from ERAI climatology
$\bar{U}$	2 m s^{-1}	Strength of turbulent velocity
α_E	1/3	Entrainment parameter
β_D	1/3	Downdraft parameter
C_d	0.001	Turbulent drag coefficient
τ_{uw}	4 h	Momentum entrainment time scale
τ_{conv}	1 h	Convective time scale
τ_D	50 days	Newtonian cooling time scale
τ_r	75 days	Rayleigh drag
ξ_s	0.5	Stratiform contribution to first baroclinic mode
ξ_c	1.25	Congestus contribution to first baroclinic mode
α_m	0.5	Ratio of convective downdraft velocity to upward mass flux velocity at top of PBL
a_0	5	Inverse convective buoyancy time scale associated with deep clouds
a_1	0.25	Relative contribution of θ_e^b to deep convection
a_2	0.75	Relative contribution of q to deep convection
a'_0	2	Inverse convective buoyancy time scale associated with congestus clouds

TABLE 2. (Continued)

Parameter	Value	Description
γ_2	0.25	Relative contribution of θ_2 to deep heating
γ'_2	0.6	Relative contribution of θ_2 to congestus heating
μ	0.2	Relative contribution of stratiform and congestus to downdrafts
α_s	0.25	Stratiform adjustment coefficient
α_c	0.1	Congestus adjustment coefficient
τ_s	3 h	Stratiform adjustment time scale
τ_c	1 h	Congestus adjustment time scale

Similarly, the climatological specific humidity values are used to obtain the moisture convergence constants $\tilde{Q}_j$, $j = 0, 1, 2$ (Khouider et al. 2011).

All datasets are from ERA-Interim (hereafter ERAI; Dee et al. 2011). The data have $0.7^\circ \times 0.7^\circ$ spatial resolution and 37 vertical levels for 3D variables. Monthly ERAI datasets are used for creating the background climatological fields, essentially for the potential and equivalent potential temperature. Other fields include the boundary layer height from the relevant variables from which the model parameters such as the cooling rates and moisture convergence constants are inferred. To estimate the surface equivalent potential temperature over the ocean, we use the skin temperature together with the corresponding saturation mixing ratio. Over the land, the skin temperature combined with the vapor mixing ratio at 2 m is used. Note that Engeln and Teixeira (2013) have used ERAI data to retrieve the PBL height based on three different methods and compared them with radiosonde data. They found that the ERAI inferred PBL height is comparable to observations in both climatology and variability, regardless of the method used.

An annual mean of this data is prepared before calculating the background fields out of it. It is worth noting here that the boundary layer height h_b used in the boundary layer equations, is also fixed to its climatological value from this dataset. The mean background is estimated from 15 years of ERAI data (January 2000 to December 2015). Daily data (i.e., θ_s and θ_{es}) are also prepared from the above 15-yr data. However, the model runs are carried out for only 10 years from 1 January 2001 to 31 December 2010.

To be more precise, the boundary layer height, 3D temperature and specific humidity fields were downloaded from the ERAI dataset for the aforementioned dates and are used to extract/calculate θ_t and θ_{et} as the values of potential temperature and equivalent potential temperature, respectively. The values of θ_t and θ_{et} are estimated at the first model level above the boundary layer, θ_b and θ_{eb} as the corresponding vertical averages from surface to the boundary layer height, and θ_m and θ_{em} are the values at 700 hPa. The 2D structures of the corresponding discrepancies used in the RCE equations above and the inferred model parameters m_0 , Q_{R1}^0 , Q_{R2}^0 , $\tilde{Q}_0$, $\tilde{Q}_1$, and $\tilde{Q}_2$ are plotted in Fig. 1.

FIG. 1. Background fields that are used in the RCE equation estimated from ERAI dataset. See [section 2](#) for the definition of the terms.

c. Maritime Continent diurnal flux

To force diurnal convective activity over the Maritime Continent, the surface fluxes in the θ_{eb} and θ_b equations (Table 1) are altered by setting the $\Delta_s\theta_e$ and $\Delta_s\theta$ discrepancies to their respective climatological means $\Delta_s\bar{\theta}_e$ and $\Delta_s\bar{\theta}$ plus a diurnally varying perturbation (Figs. 1a,d). We note that the MCM with the boundary layer dynamics, is used here as a convective parameterization, produces a realistic diurnal cycle of convection both over land and over the ocean, when tested with a similarly varying diurnal flux in a single column setting (i.e., without spatial variation; Frenkel et al. 2011a,b). For simplicity, only the $\Delta_s\theta_e$ term was perturbed.

It is found that in the case of the diurnal cycle over wetland surfaces and over the ocean, the sensible heat forcing has a negligible effect on the MCM response due to weak Bowen ratios (Frenkel et al. 2011b).

To mimic the unorganized and multiscale nature of MC diurnal convection, we use a formula for the surface fluxes that bears both stochastic and deterministic parts. We set

$$\Delta_s\theta_e(x_{isl}) = \gamma \left[\sin\left(\frac{t2\pi}{\text{day}}\right) \right]^+,$$

$$\gamma = -\bar{\beta} \ln(\mu) + \beta_{\min}. \quad (6)$$

TABLE 3. Sensitivity experiments varying the strength of the diurnal flux over the MC. The parameters $\alpha_s = 0.25$, $\alpha_c = 0.1$, $\tau_{\text{conv}} = 1$ h, $\alpha_E = 1/3$, $\beta_D = 1/3$, $a_1 = 0.25$, $a_2 = 0.75$, and $a_0 = 5$ are fixed for all experiments.

Cases	Island flux	Mean IFlux strength	Flux type
Case 0	None	0	
Case 1	$\bar{\beta} = 7, \beta_{\min} = 2$	9	Stochastic
Case 2	$\bar{\beta} = 10, \beta_{\min} = 5$	15	
Case 3	$\bar{\beta} = 15, \beta_{\min} = 5$	20	
Case 4	$\bar{\beta} = 20, \beta_{\min} = 10$	30	
Case 5	$\bar{\beta} = 30, \beta_{\min} = 10$	40	
Case 7	$\bar{\beta} = 0, \beta_{\min} = 10$	10	Deterministic
Case 8	$\bar{\beta} = 0, \beta_{\min} = 20$	20	
Case 9	$\bar{\beta} = 0, \beta_{\min} = 30$	30	
Case 10	$\bar{\beta} = 0, \beta_{\min} = 40$	40	
Case 11	$\bar{\beta} = 0, \beta_{\min} = 50$	50	

Here, x_{isl} represents all grid points that are on the MC islands, $\bar{\beta}$ is the mean Maritime Continent flux strength, and $\beta_{\min}$ is the minimum strength while μ is a random number uniformly chosen between 0 and 1 so that γ is exponentially distributed with mean $\bar{\beta} + \beta_{\min}$. Both $\bar{\beta}$ and $\beta_{\min}$ are tunable parameters that set the contributions of the stochastic and deterministic parts to the total strength of the diurnal surface fluxes (see Table 3). $\bar{\beta}$ is based on the mean value of background $\Delta\theta_e^e$ (Fig. 1a). Figure 2 summarizes the overall structure of the

imposed diurnal surface flux, when $\bar{\beta} = 20$ K and $\beta_{\min} = 10$ K. The Maritime Continent is defined as the region encompassing 15°S–12°N, 90°–170°E (Fig. 2d).

d. Experiment design

The HOMME-MCM model with boundary layer dynamics described above is ran for 10 years with prescribed SSTs from 1 January 2001 to 31 December 2010. Apart from the MC islands, the model is forced by surface fluxes (estimated from ERAI skin surface temperatures) provided, every 24 h, as described below, and the outputs were saved daily. With the surface pressure is set to 1000 hPa everywhere, θ_e [in Eq. (6)] is calculated separately for land and ocean as follows. Over the land, the skin temperature (ST) and mixing ratio (r) at 2 m height are used to calculate the equivalent potential temperature as

$$\theta_e = \left(\text{ST} + \frac{L_v}{C_p} r \right) \left(\frac{P_0}{p} \right)^\kappa, \quad (7)$$

where $P_0 = 1000$ hPa, p is the surface pressure and $\kappa = R_d/C_p$ is the ratio of the gas constant over the heat capacity at a constant pressure of dry air. Over the ocean, we use the SST and the associated saturated mixing ratio (r^*):

$$\theta_e = \left(\text{SST} + \frac{L_v}{C_p} r^* \right) \left(\frac{P_0}{p} \right)^\kappa, \quad (8)$$

FIG. 2. (a) Diurnal profile, (b) island flux (K), and (c) histogram of the island flux applied over the MC. (d) MC continent domain (15°S–12°N, 90°–170°E) where the diurnal island flux is applied.

FIG. 3. Wave–frequency spectra of the symmetric component of precipitation (estimated over 5°S–5°N) constructed from the outputs of the model sensitivity experiments for the stochastic case (see Table 3). While cases 1–5 represent stochastic diurnal forcing, case 0 represents no imposed diurnal forcing.

FIG. 4. As in Fig. 3, but for the deterministic experiments (see Table 3).

FIG. 5. Maps of the unfiltered variance of precipitation estimated from the outputs of the model sensitivity experiments for the stochastic case. Note that case 0 represent no imposed diurnal forcing.

$$r^* = 0.622 \frac{e_s}{p}, \quad (9)$$

$$e_s(\text{SST}) = e_{s0} \exp \left\{ \left[\frac{L_v(\text{SST})}{R_v} \right] \left(\frac{1}{T_0} - \frac{1}{\text{SST}} \right) \right\}, \quad (10)$$

$$\approx 611 \exp \left[\frac{17.67(\text{SST} - T_0)}{\text{SST} - 29.65} \right], \quad (11)$$

where T_0 is the triple-point temperature of water = 273.16 K

To explore the parameter space and to understand the impact of the MC diurnal convection on the intraseasonal variability, we conduct a set of sensitivity experiments where the key variable [γ , Eq. (6)], which controls the strength of the diurnal flux over MC is systematically increased in both its stochastic and deterministic contributions from small to large values. In particular, experiments with $\bar{\beta} = 0$ is fully deterministic and are systematically differentiated from those with $\bar{\beta} > 0$ that is called stochastic. A total of five experiments varying the strength of the island flux is carried out for each stochastic and deterministic cases (Table 3). The goal is to identify the relative changes in the mean features of MJO; especially its propagation characteristics when the MC diurnal flux is changed. We also conducted an experiment with no island flux (case 0) for comparison.

3. Results

a. Power spectra of precipitation

Although we carried out the simulations for all seasons, the analysis is restricted to the last 10 winter seasons to focus on the Northern Hemisphere intraseasonal variability, which is known to be affected by the MC barrier effect. The formation and subsequent eastward propagation of MJO disturbances are more prominent in the winter season (Zhang 2005). The characteristics of the simulated MJO in various model experiments can be best illustrated with the help of the wavenumber–frequency [Wheeler–Kiladis (WK)] diagrams (Wheeler and Kiladis 1999). The WK diagrams of the five sensitivity experiments associated with the stochastic case (Table 3) are illustrated in Fig. 3. The model simulates the MJO power at zonal wavenumbers 2 and 3 without altering the island surface fluxes (IFlux; case 0 in Fig. 3a). The model has a systematic bias in simulating the correct Kelvin wave speed ($\approx 15 \text{ m s}^{-1}$). The Kelvin waves simulated by the model is around 22 m s^{-1} (Fig. 3a). When the intensity of IFlux increases, the power of the Kelvin waves decreases (see Figs. 3b–f). Although the MJO signal is visible in all the cases, the Kelvin wave signal weakens significantly when IFlux becomes very large (case 4 and case 5). A more or less similar pattern is simulated by the model in the deterministic IFlux experiments (Figs. 4a–f).

FIG. 6. As in Fig. 5, but for the deterministic experiments (see Table 3).

However, the MJO signal persists even at the very large IFlux values (Figs. 4d–f), where the imposed maximum strength is comparable to the maximum mean strength in stochastic cases 4 and 5, as can be surmised from Table 3. Note that since the large-scale pattern of MJO envelope depends on several other factors like boundary forcing and convection schemes, the change in MJO characteristics are only limited to finescale features. Also, the strength of the Kelvin wave weakens only in the very high IFlux forcing (case 9, case 10, and case 11). The Kelvin wave power of precipitation decreases as the IFlux increases in both stochastic and deterministic forcing experiments. However, the weakening of Kelvin wave power of precipitation happens much earlier in the stochastic case compared to the deterministic forcing. As elucidated below, this is not the case for the Kelvin wave power of zonal winds, which suggests that only convectively coupled Kelvin waves (CCKWs) are affected in this fashion by the island diurnal convection. The contrasting behavior of Kelvin wave characteristics in precipitation and zonal winds may be a result of the weakening of the MJO itself as these waves are known to be often embedded with the MJO envelope over this region of the globe (Kiladis et al. 2009; Khouider et al. 2012).

b. Intraseasonal variance of precipitation

A simple and straightforward way to depict the contrasting changes in intraseasonal variability simulated by the various

sensitivity experiments (Table 3) is by estimating and comparing the associated variances. Figure 5 shows the unfiltered variance of precipitation simulated by the experiments with stochastic IFlux forcing. In general, the precipitation variance over the Indo-Pacific warm pool decreases with the increase in IFlux forcing. The variance averaged over the Indo-Pacific warm pool (as indicated by the red box in Fig. 5a) is marked in the top-left corner for a quantitative comparison between the experiments. As the island flux increases, the precipitation gets blocked over the MC as indicated by the increased variance there (over MC, 10°S–0°, 120°–160°E) while the variance over the entire tropics weakens (Figs. 5a–f). The weakening of precipitation variance is more pronounced over the west Pacific, which is believed to be associated with the MC IFlux (as indicated by the blue box in Fig. 5a) that blocks the propagation of precipitation disturbances through the MC. A similar decreasing variance in precipitation with the increase in IFlux forcing is evident in the deterministic case as well (Figs. 6a–f).

To further delineate which part of the precipitation spectrum is most affected by this MC “barrier effect,” we plot in Figs. 7 and 8 the bandpass filtered (20–100 days) precipitation anomalies for the stochastic and deterministic experiments. When the strength of the MC diurnal flux increases, the intraseasonal variance over the MC/west Pacific (10°S–0°, 120°–160°E) increases. This indicates the critical role of MC diurnal forcing

FIG. 7. Maps of the bandpass filtered (20–100 days) variance of precipitation estimated from the outputs of the model sensitivity experiments for the stochastic case. Note that case 0 represents no imposed diurnal forcing.

in blocking the propagation of MJOs across the MC. This blocking is further illustrated in Fig. 9, where the lag–longitude plots of intraseasonally filtered precipitation anomalies are plotted. The simulation without any MC diurnal IFlux shows realistic eastward propagation of MJO anomalies (Fig. 9a) compared to observations (Waliser et al. 2009), consistent with previous studies using the HOMME-MCM models (Ajayamohan et al. 2013; Deng et al. 2015). Moreover, the MJO propagates a great distance across and beyond the MC in this case without diurnal forcing (Fig. 9a). Besides, MJO propagation weakens as it passes the MC boundary (indicated by the dashed line at $\sim 90^\circ\text{E}$). However, when the diurnal IFlux forcing is applied over the MC, the eastward propagation of MJO anomalies appears to be abruptly blocked and weakened over the MC (Fig. 9b). The blocking intensifies as the strength of the IFlux forcing increases (Fig. 9c). The barrier effect over the MC seems to be more efficient when a stochastic diurnal forcing (Figs. 9d–f) is enforced compared to the deterministic forcing. Similar to deterministic IFlux forcing, the blocking effect enhances as the strength of the stochastic IFlux forcing increases.

As can be seen from these figures (Figs. 7–9), the impact of the increased MC diurnal flux on the intraseasonal variance is more dramatic, and indeed it is more so over the western Pacific part of the warm pool region. This further confirms the

“barrier effect” that the diurnal cycle of convection has on the propagation of intraseasonal precipitation disturbances associated with the MJO past the MC (Kerns and Chen 2016; Ling et al. 2019). This point is further illustrated by analyzing the dynamical variables.

c. Impact on dynamical variables and the local climatology

The relationship between IFlux strength and variances over the west Pacific warm pool region is further illustrated in Fig. 10 as a scatterplot. Both unfiltered (Figs. 10a–c) and intraseasonally filtered (Figs. 10d–f) variances of three variables, viz., precipitation, 850 hPa zonal wind, and 200 hPa zonal winds, are used for this calculation. Each data point in these panels indicates a single experiment (from Table 3 and Figs. 5–8), reported on the x axis as the IFlux diurnal convection strength, which increases from 0 for case 0 to 40 for case 5.

A near-linear relationship between the strength of the IFlux and unfiltered/intraseasonal precipitation variance over the west Pacific is not evident (Figs. 10a,d). However, the strength of the precipitation variance decreases for the first three cases (cases 0–3). Case 3 and case 4 show a mild or insignificant increase in precipitation variance with the increase of IFlux, indicating that a stagnation level with regard to the strength of IFlux is reached. Note that the bandpass filtered variance also

FIG. 8. As in Fig. 7, but for the deterministic experiments (see Table 3).

shows a similar relationship between the strength of IFlux and precipitation variance as in unfiltered case (Figs. 10a,d). In other words, the relationship between the strength of MC IFlux and precipitation variance remains the same in the intra-seasonal and the unfiltered bands. The relationship between the variance and the strength of IFlux is much more linear in the dynamic variables like lower-level zonal winds (Figs. 10b,e). The variance of low-level zonal winds decreases with the strength of the IFlux over the western Pacific warm pool. Like in precipitation, the decrease in variance is more evident for experiments 0–3 and then appears to remain stagnant with the increase in IFlux. The variance of upper-level winds, on the other hand, increases with the strength of the MC IFlux and shows a near-linear relationship (Figs. 10c,f). The association between the strength of the IFlux and variance of the three fields considered (precipitation and low-level and upper-level zonal winds) for the deterministic case (Figs. 11a,f) is almost similar to that of the stochastic case. However, the strength of the IFlux is nearly double the strength of the stochastic case (cf. x axis in Figs. 10 and 11) for comparable variances (cf. y axis in Figs. 10 and 11). In a sense, the stochasticity is more detrimental in blocking the propagation of the MJO over the MC even when the mean diurnal fluxes are reasonably low. This is perhaps an indication why GCMs that cannot represent organized convection over the islands, from which the diurnal convection appears as noisy, may have a harder time overcoming

the MC barrier. These results hint at why the convective parameterization schemes that better represent organized features of convection (Biello et al. 2010; Peters et al. 2017) perform better (Deng et al. 2015; Goswami et al. 2017; Ma et al. 2019).

An intriguing feature in Figs. 10 and 11 is the near-linear increase of both the filtered and unfiltered variances of the 200 hPa zonal wind. To shed some light on this behavior, we plot the power spectrum WK diagrams of the upper-level winds for the five deterministic experiments (Table 3) compared with the case without imposed diurnal forcing (Fig. 12). As we can see, the increase in the island diurnal flux results in an increase of Kelvin waves contrary to WK diagrams of precipitation (cf. to Fig. 4). Those waves appear to line up with the 50 m equivalent depth dispersion relation, corresponding to second baroclinic dry Kelvin waves (Khouider 2019). This is consistent with an earlier study by Ajayamohan et al. (2013), who found that the large-scale flow patterns are associated with planetary-scale dry Kelvin waves that are triggered by preceding MJO events and circumnavigate the globe. The increased convection over the MC (see Figs. 13d–f) due to the increased island flux seems to trigger stronger and more of these waves.

We have discussed the behavior of the model in response to the strength of the diurnal flux over the MC. To gain some insight and elucidate the possible dynamical reason behind this dramatic change in variance in response to the changes in the

FIG. 9. Lag-longitude plots of bandpass filtered precipitation averaged over the equatorial belt (10°S – 10°N) from selected experiments (see Table 3). (a) Case 0 (no IFlux over MC), (b),(c) case 7 and case 11 (deterministic IFlux), and (d)–(f) case 1, case 3, and case 5 (stochastic IFlux). Lag correlation is estimated using a reference time series averaged over 10°S – 5°N , 75° – 100°E and correlated over all longitudes. The stippled black line indicates the western boundary of MC (95°E) where the diurnal IFlux forcing is applied in the model experiments.

strength of the IFlux forcing, the effect of this forcing on the local circulation climatology (see Fig. 1 of Jourdain et al. 2013) is examined. Figure 13 shows the north–south overturning circulation (regional Hadley circulation) averaged over the MC region, as indicated by the black box in Fig. 2d, for the five deterministic experiments (Table 3). Figure 13a (case 0) corresponds to the default case, with no additional diurnal flux over the MC islands. Two ascending cells over the MC and southern IO appear to converge over the equatorial IO in this case. This convection is stronger over the channels north and south of the archipelago and more so in the north according to the morphology of the MJO convection as it propagates past the MC (Zhang and Ling 2017). This behavior persists for weaker diurnal island fluxes as is the case for case 7 but as the descending cell over the equatorial region weakens considerably (Fig. 13b). When the strength of the MC IFlux increases further, the descending cell over the equatorial region is replaced by an upward motion in response to the dominant IFlux forcing. In short, the regional Hadley circulation reverses as the strength of the IFlux forcing increases (Figs. 13a–f). The overturning circulation extends up to about 400 hPa in all cases. Although we are prescribing the diurnal cycle only, the changes in the regional Hadley circulation is evident until 400 hPa. Note that there is no cloud radiative feedback in the model.

The MJO events that propagate across the MC rain much more over the sea than over the land compared to the events that are blocked (Zhang and Ling 2017). The strong island convection, subsequent increased soil moisture and reduced diurnal amplitude of land convection weakens the organized convection over the surrounding waters (Ling et al. 2019). In other words, the diurnal cycle in land convection acts as an intrinsic barrier for smooth propagation of MJOs across the MC. The propagation of MJOs across MC can also be modulated by the background moisture anomalies or the mean moisture gradient (DeMott et al. 2018; Ahn et al. 2020). The horizontal advection of the moisture by the wind anomalies is another factor that determines the propagation characteristics of MJO over the MC (Ahn et al. 2020). When juxtaposed with these studies, our results on the reversal of the local mean circulation indicate that the strong island convection weakens the organized convection over the surrounding waters. This leads to the stalling of the MJO propagation across the MC.

4. Conclusions

As elucidated in section 1, the Maritime Continent acts as a barrier for the smooth eastward propagation of MJO disturbances. The character of the MJO propagation undergoes

FIG. 10. Relationship between the strength of the island flux and the western Pacific warm pool (as indicated by the blue box in Figs. 5a, 6a, 7a, and 8a) precipitation, low-level and upper-level zonal wind variances for the stochastic experiments (Table 3). (a)–(c) The total (unfiltered) variance and (d)–(f) the intraseasonally filtered variance. See text for details.

notable changes while passing through the MC in the sense that either the MJO stalls over the MC or the propagation speed weakens. The strong diurnal variability over the MC is often cited as a major factor that controls the behavior of MJO over the MC. Several peculiar features of MC like persistent high SST over the Indo-Pacific warm pool, unique geography, and deep convective activity support an unstable environment that favors modification of MJO disturbances. Here, we study the relative contribution of the strength of the diurnal variability in reshaping the propagation characteristics of MJO with the help of a coarse-resolution AGCM. Specifically, we examine how a simulated variability of the diurnal cycle influences the MJO propagation. While the simulated variability is not realistic, it can be viewed as a proxy for both regular and chaotic representation of the

diurnal convection by state-of-the-art climate models that are often at odds with the reality (Yang and Slingo 2001; Neale and Slingo 2003; Klingaman et al. 2017; Martin et al. 2017; Peters et al. 2017).

We ran a coarse-resolution HOMME model as a dry dynamical core coupled to the multcloud parameterization. The ability of the aquaplanet version of this model in simulating convectively coupled waves and intraseasonal oscillations like MJO has been previously demonstrated and widely documented (Khouider et al. 2011; Ajayamohan et al. 2013, 2014, 2016; Deng et al. 2015, 2016). Here, we coupled the model with the dynamical bulk boundary layer model (Waite and Khouider 2009; Stevens 2006; Frenkel et al. 2011a; see section 2) to study the influence of diurnal variability on MJO propagation. The effect of deterministic and stochastic diurnal forcings are compared to assess the

FIG. 11. As in Fig. 10, but for the deterministic experiments (see Table 3). An additional model experiment ($\bar{\beta} = 0$, $\beta_{\min} = 60$) is also included in this plot.

impact of stochasticity generated convection of the MC diurnal convection on MJO propagation. A set of sensitivity experiments is conducted by systematically varying the strength of the MC diurnal flux. Note that five such experiments were conducted for both stochastic and deterministic cases (Table 3). The intention is to identify the relative changes in the mean features of MJO, and mostly its propagation characteristics when the MC diurnal flux is altered. These experiments were compared with an experiment with no island flux (case 0) as a reference. The HOMME-MCM model with boundary layer dynamics described above is ran for 10 years with prescribed SSTs from 1 January 2001 to 31 December 2010. The analysis is restricted to 10 winter seasons, where the MJO disturbances are more prominent. Lower-level, upper-level zonal winds and precipitation fields from the model simulations are analyzed.

It is found that, in general, the precipitation variance over the Indo-Pacific warm pool decreases with the increase in IFlux

forcing. This suggests the blocking of precipitation over MC as indicated by the increased variance there while the variance over the entire tropics weakens (Figs. 5 and 6). The increased diurnal convection over the MC induces an increase in precipitation variance over the MC while the associated weakening of precipitation variance is more pronounced over the west Pacific. The “barrier effect” that the diurnal cycle of convection has on the propagation of MJO is demonstrated by the severely weakened 20–100 day intraseasonal variance over the west Pacific (Figs. 7–9).

In summary, the relationship between the strength of MC IFlux and precipitation variance remains the same in both the intraseasonal and unfiltered regimes. The variance of low-level zonal winds and precipitation decreases with the strength of the IFlux over the western Pacific warm pool (Figs. 10b,e and 11b,e). On the other hand, the variance of upper-level winds increases with the strength of the MC IFlux and shows a near-linear

FIG. 12. Wavenumber–frequency spectra of the symmetric component of upper-level (200 hPa) zonal winds constructed from the outputs of the six model sensitivity experiments. While cases 1–5 represent deterministic diurnal forcing, case 0 represent no imposed diurnal forcing.

FIG. 13. Regional Hadley cell for the deterministic experiments. Meridional winds and vertical velocity (vectors; m s^{-1}) and total heating (shaded, QHe; K day^{-1}) averaged over $70^{\circ}\text{--}130^{\circ}\text{E}$ and averaged over a 10 yr period. Note that case 0 represent no imposed diurnal forcing.

relationship (Figs. 10c,f and 11c,f). It appears that the upper-level winds are associated with dry Kelvin waves of 50 m equivalent depth contrary to precipitation and lower-level winds (Fig. 12). The large-scale flow patterns are associated with planetary-scale dry Kelvin waves that are triggered by preceding MJO events and circumnavigate the globe (Ajayamohan et al. 2013).

The notable difference between deterministic and stochastic cases resides in their relative strengths in IFlux needed for MC blocking. The strength needed to attain the same level of blocking for the deterministic case is almost double to that of the stochastic case (Figs. 10 and 11). The stochastic case seems to favor stronger blocking of MJO propagation, especially when the diurnal flux is low. This authenticates the fact that the models, which cannot represent organized convection associated with the diurnal cycle over the MC, struggles in overcoming the MC barrier, resulting in poor skill in simulating MJO propagation. It is further found that the regional Hadley

circulation reverses as the strength of the IFlux forcing increases (Fig. 13), suggesting a dynamical explanation behind the dramatic changes in variance in response to the changes in the strength of the IFlux forcing and the MC barrier effect in general. The strong upward motion on the equator due to the increased MC convective heating is perhaps detrimental for the MJO dynamics and wave propagation in general.

While this study provides some guidelines on how the MC diurnal convection can indeed inhibit the propagation of the MJO beyond the MC by reducing the intraseasonal variance over the western Pacific, it does not take into account the realistic day-to-day variability of organized convection over the MC. However, the present model setting can easily be extended by conducting sensitivity experiments with observed surface flux variability. For example, we can consider the effect of mesoscale convective systems with time scales of 1 to 2 days that are often observed to propagate back and forth across the coastal boundaries of the MC and introduce the important day

to day variability of the diurnal convective signal (Moncrieff 2013; Bergemann et al. 2015).

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