Effects of Rotation on the Multiscale Organization of Convection in a Global 2D Cloud-Resolving Model

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ABSTRACT

Atmospheric convection exhibits distinct spatiotemporal variability at different latitudes. A good understanding of the effects of rotation on the multiscale organization of convection from the mesoscale to synoptic scale to planetary scale is still lacking. Here cloud-resolving simulations with fixed surface fluxes and radiative cooling are implemented with constant rotation in a two-dimensional (2D) planetary domain to simulate multiscale organization of convection from the tropics to midlatitudes. All scenarios are divided into three rotation regimes (weak, order-one, and strong) to represent the idealized ITCZ region (0° – 6° N), the Indian monsoon region (6° – 20° N), and the midlatitude region (20° – 45° N), respectively. In each rotation regime, a multiscale asymptotic model is derived systematically and used as a diagnostic framework for energy budget analysis. The results show that planetary-scale organization of convection only arises in the weak rotation regime, while synoptic-scale organization dominates (vanishes) in the order-one (strong) rotation regime. The depletion of planetary-scale organization of convection as the magnitude of rotation increases is attributed to the reduced planetary kinetic energy of zonal winds, mainly due to the decreasing acceleration effect by eddy zonal momentum transfer from mesoscale convective systems (MCSs) and the increasing deceleration effect by the Coriolis force. Similarly, the maintenance of synoptic-scale organization is related to the acceleration effect by MCSs. Such decreasing acceleration effects by MCSs on both planetary and synoptic scales are further attributed to less favorable conditions for convection provided by weaker background vertical shear of the zonal winds, resulting from the increasing magnitude of rotation.

1. Introduction

Atmospheric convection plays a crucial role in the horizontal and vertical transports of momentum, heat, and moisture of large-scale circulation of Earth (Schneider 2006). After decades of observational studies based on satellite and in situ measurements, it is apparent now that the spatiotemporal variability of convection has distinct characteristics at different latitudes (Riemann-Campe et al. 2009). Specifically, tropical convection is organized in a hierarchy of spatiotemporal scales, ranging from a cumulus cloud of

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several kilometers and a few minutes to MCSs (Houze 2004) of several hundred kilometers and a few hours to convective coupled equatorial waves (CCEWs) (Kiladis et al. 2009) of thousands of kilometers and 1–2 weeks to the Madden–Julian oscillations (MJOs) (Zhang 2005) of 10 000 km and 1–3 months. In contrast, convection in the subtropics is dominated by synoptic-scale convective disturbances such as low pressure systems in the Indian monsoon trough region (Hurley and Boos 2015). Theoretically, the magnitude of rotation can dramatically influence the behavior of geophysical flows (Majda 2000). In the midlatitudes, the strong rotation leads to a strict temporal frequency scale separation between potential vorticity dynamics and fast gravity waves. In contrast, the weak rotation in the tropics does not induce a time-scale separation any more but allows multiscale organization of convection in the presence of warm surface temperature and abundant moisture (Majda 2012).

Contemporary global climate models (GCMs) struggle to accurately simulate the multiscale organization of tropical convection. In fact, present-day GCMs still have difficulty in simulating key features of propagating MJOs (Jiang et al. 2015), although predictions of the MJO have improved over the past decade (Kim et al. 2018). Furthermore, it is observed that the MJO is a slowly eastward-moving planetary-scale envelope that contains a few superclusters of cloudiness with numerous embedded cloud clusters (Nakazawa 1988; Chen et al. 1996). Even good GCMs fail to satisfyingly simulate these multiscale features (Guo et al. 2015). It is hypothesized here that the poorly simulated MJOs in the GCMs are due to an inadequate treatment of multiscale interactions of convection, especially the upscale impact of organized tropical convection such as MCSs that are poorly resolved in the coarse-resolution GCM simulations.

To address this issue, it is necessary to obtain a better understanding of spatiotemporal scale selection and multiscale interactions of convection. With the development of computational resource, cloud-resolving models (CRMs) have become a practically useful tool for simulating organized convection in a fine horizontal resolution of a few kilometers (Khairoutdinov and Randall 2003; Miura et al. 2007; Tao and Moncrieff 2009; Guichard and Couvreux 2017). In particular, the 2D CRM simulations provide a cheap way to study the multiscale organization of convection in a planetary domain. For example, the idealized 2D CRM simulation by Grabowski and Moncrieff (2001) showed that convection in background easterly winds is organized in a two-scale structure with a synoptic-scale envelope moving eastward and numerous embedded MCSs moving westward. Slawinska et al. (2014)

showed that the Walker circulation over a warm pool exhibits intraseasonal variability with outward (inward)-moving synoptic-scale systems during its expansion (contraction) phases. Due to expensive computational cost, many three-dimensional (3D) CRM simulations only focused on radiative–convective equilibrium in small domains (Held et al. 1993; Bretherton et al. 2005). In the absence of rotation, those disordered and scattered small-scale clouds arising from initial disturbances in a moist unstable environment coalesce into large-scale patches of convection, which is known as self-aggregation (Bretherton et al. 2005; Muller and Held 2012; Wing and Emanuel 2014). Bretherton et al. (2005) recognized the self-aggregation as an instability driven by convection–water vapor–radiation–surface fluxes feedbacks. However, those theories for explaining large-scale organization of convection mostly focus on thermodynamic effects, while dynamic effects due to multiscale interactions are overlooked. Moreover, the absence of rotation makes the model setup less realistic.

In fact, several studies have been conducted to investigate the effects of rotation on scale selection and multiscale organization of convection. Majda et al. (2015) used the multicloud model (Khouider and Majda 2006a,b,c, 2007) with either a deterministic (Khouider and Majda 2008a,b) or stochastic (Khouider et al. 2010; Deng et al. 2015; Goswami et al. 2017) convective heating closure to simulate organized convection in a rotating 2D flow. They concluded that the planetary rotation is one of important players in the diminishing of organized convection and convectively coupled gravity wave activity, and deep convection activity in the stochastic model simulations becomes patchy and unorganized in the subtropics and midlatitudes. The 2D nonhydrostatic anelastic model simulation by Liu and Moncrieff (2004) indicated that rotation-induced localized descent stabilizes and dries the neighborhood of convective region, explaining the fact that the tropics is a preferred region for convective clustering. In general, planetary rotation has significant impact on background sounding of thermodynamic fields and vertical shear, the latter of which plays a crucial role in promoting organized convection (Newton and Rodebush Newton 1959; Moncrieff 1981; Moncrieff and Liu 1999; Tompkins 2001).

The goals of this paper include the following four aspects: 1) using a global 2D CRM to simulate multiscale organization of convection in three regimes with weak, order-one, and strong rotation; 2) deriving a multiscale asymptotic model for upscale and downscale impacts in each rotation regime and using it as a diagnostic framework for energy budget analysis; 3) explaining why planetary-scale organization diminishes in the weak rotation regime as the magnitude of rotation increases and

investigating the role of eddy transfer of momentum, temperature, and equivalent potential temperature from meso- and synoptic-scale fluctuations; and 4) explaining why synoptic-scale organization persists in the order-one rotation regime but diminishes in the strong rotation regime.

Here we use the System for Atmospheric Modeling (SAM) (Khairoutdinov and Randall 2003) to investigate the effects of rotation on the multiscale organization of convection. In particular, Brenowitz et al. (2018) configured the model in a global 2D periodic domain to simulate organized convection without the rotation. With both radiative cooling and surface fluxes fixed, the simulation in background easterly winds still produces an eastward-moving planetary-scale envelope of convection with multiple superclusters of cloudiness and numerous embedded clusters. To identify physical mechanisms behind the multiscale organization, Brenowitz et al. (2018) decomposed the model outputs into meso-, synoptic-, and planetary-scale components and concluded the key role of multiscale interactions in promoting large-scale organization of convection based on energy budget analysis. Here we configure the SAM in a similar way as Brenowitz et al. (2018) but with the Coriolis force. The magnitude of rotation is varied to represent three different regimes, including the ITCZ regime with weak rotation, the Indian monsoon trough regime with order-one rotation, and the midlatitude regime with strong rotation. Considering the fact that convective activity in both the ITCZ and the Indian monsoon trough is organized in an elongated strip, it is believed here that the 2D configuration is a physically suitable and computationally cheap way for modeling real physical processes.

In each regime, we derive a multiscale model by following the multiscale asymptotic methods (Majda and Klein 2003; Majda 2007) and use it as a diagnostic framework for energy budget analysis. In particular, the multiscale models in the weak and order-one rotation regimes are derived under the standard physical scaling in the tropics (Majda 2007). Consequently, the governing equations across synoptic- and mesoscales are similar to the mesoscale equatorial synoptic dynamics (MESD) model (Majda 2007), and those across planetary- and synoptic-scales resemble the intraseasonal multiscale moist dynamics (IMMD) model (Biello and Majda 2010; Back and Biello 2018). Notably, the MESD model has been used to study the upscale impact of MCSs on convectively coupled Kelvin waves (CCKWs) (Yang and Majda 2017, 2018) and 2-day waves (Yang and Majda 2019). In contrast, the multiscale model in the strong rotation regime follows the classic quasigeostrophic (QG) scaling (Vallis 2017).

We run 10 SAM simulations under the similar configuration as Brenowitz et al. (2018) but with increasing magnitude of rotation. Several key results about the effects of rotation are obtained. First of all, planetary-scale organization of convection only arises in the weak rotation regime, while synoptic-scale organization persists in the order-one rotation regime but diminishes as the magnitude of rotation further increases. As summarized by the schematic diagram in Fig. 9, the diminishment of planetary-scale organization is attributed to two changing effects in terms of planetary kinetic energy budget of zonal winds, including decreasing acceleration effect by eddy zonal momentum transfer from mesoscale fluctuations and increasing deceleration effect by the Coriolis force. As for the acceleration effect from upscale impact of MCSs, its decreasing strength is attributed to less favorable conditions for convection provided by weaker background vertical shear of zonal winds, resulting from the increasing magnitude of rotation. Similarly, the maintenance of synoptic-scale organization in the order-one rotation regime and its diminishment in strong rotation regime are also related to the decreasing acceleration effect from upscale impact of MCSs, as summarized by the schematic diagram in Fig. 15.

The rest of the paper is organized as follows. Section 2 describes the model configuration and experiment design. Section 3 shows the spatiotemporal variability of brightness temperature and the zonal-mean climatology of winds and thermodynamic fields with different magnitude of rotation. A multiscale decomposition method is introduced to decompose total fields into domain-mean and planetary-, synoptic-, and mesoscale fluctuations. Section 4 investigates the planetary-scale kinetic energy budget of zonal and meridional winds and available potential energy in the weak rotation regime, and highlights the key role of eddy transfer of momentum, temperature, and equivalent potential temperature. Section 5 does a similar energy budget analysis for synoptic-scale flow fields in the order-one rotation regime, while section 6 considers the strong rotation regime. The paper concludes with a discussion in section 7.

2. Model configuration and experiment design

Thanks to its easy configuration and fast execution, the SAM has been used widely to simulate large-scale organization of convection in idealized domain geometry (Bretherton et al. 2005; Wing and Emanuel 2014; Bretherton and Khairoutdinov 2015; Wing and Cronin 2016). The SAM, version 6.11.1, is used here under the similar configuration as the QSTRAT simulation

TABLE 1. Coriolis force parameter [$f = 2\Omega \sin(\phi)$] and the corresponding time scale ($1/f$) in the 10 cases.

Regime	Latitude ($^{\circ}\text{N}$)	f (s^{-1})	$1/f$ (h)
No rotation	0	0	∞
Weak rotation	1	2.5×10^{-6}	109.1
	3	7.6×10^{-6}	36.4
	5	1.3×10^{-5}	21.9
Order-one rotation	9	2.3×10^{-5}	12.2
	14	3.5×10^{-5}	7.9
	20	5.0×10^{-5}	5.6
Strong rotation	27	6.6×10^{-5}	4.2
	35	8.4×10^{-5}	3.3
	45	1.0×10^{-4}	2.7

in [Brenowitz et al. \(2018\)](#) but with the Coriolis force. All simulations use the single-moment microphysics, Smolarkiewicz's MPDATA advection scheme with monotonic corrector, and the 1.5-order closure (prognostic SGS turbulent kinetic energy) subgrid-scale scheme. To exclude effects of surface fluxes, we perform all simulations over a uniform 300.15-K sea surface temperature (SST) ocean surface with latent

and sensible heat fluxes fixed at 210.6 and 31.20 W m^{-2} , respectively. To avoid effects of active radiation, we prescribe a fixed radiative cooling of 1.5 K day^{-1} below 150 hPa and a constant stratospheric heating of 4.5 K day^{-1} above. The stratospheric heating increases stratification of the atmosphere near the tropopause, turning the troposphere into a rigid-lid scenario. Similar to [Grabowski and Moncrieff \(2001\)](#), the zonal winds are nudging toward -10 m s^{-1} easterly background winds with nudging time scale of 1 day. A sponge layer is added in the upper one-third of the model domain to damp gravity waves. The 2D planetary domain has $2^{15} = 32\,768$ -km zonal extent in a 2-km horizontal grid spacing and 27-km vertical extent with 64 vertical levels. All simulations are run for 100 days, and the last 80-day solutions are used for diagnostic analysis.

Here we repeat the nonrotating simulation in [Brenowitz et al. \(2018\)](#) as the control experiment and run another nine simulations with increasing magnitude of rotation from the tropics to the midlatitude in the Northern Hemisphere (NH). The counterparts in the Southern Hemisphere can be inferred based on the mirror symmetry about the equator.

TABLE 2. Multiscale asymptotic model across planetary, synoptic, and mesoscales in the weak rotation regime [regime 1; $\hat{f}$ from $\mathcal{O}(\varepsilon)$].

Space and time scales	Governing equations	Variables
Trade winds (planetary/intraseasonal)	$DU/DT = -P_X - \hat{d}(U - U_0)$ $D\Theta/DT + N^2 W = -\hat{d}_\theta \Theta + S_\theta$ $P_z = \Theta$ $U_X + \rho_0^{-1}(\rho_0 W)_z = 0$ $(DQ/DT) - Q_0 = -S_\theta$	$\hat{d}, \hat{d}_\theta \text{ from } \mathcal{O}(\varepsilon)$ $D/DT = \partial/\partial T + U(\partial/\partial X) + W(\partial/\partial z)$ $U, P, \Theta, Q \text{ from } \mathcal{O}(1)$ $V, W, S_\theta \text{ from } \mathcal{O}(\varepsilon)$ $\rho_0 = \rho_0(z)$ $N^2 = 1$
Planetary/intraseasonal anomalies from the climatology	$Du/DT + uU_X + wU_z - \hat{f}V = -p_X - \hat{d}u$ $-\rho_0^{-1}(\rho_0 \langle w^* v^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' u' \rangle^p)_z$ $DV/DT + \hat{f}u = -\hat{d}V$ $-\rho_0^{-1}(\rho_0 \langle w^* v^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' v' \rangle^p)_z$ $D\theta/DT + u\Theta_X + w\Theta_z + N^2 w = -\hat{d}_\theta \theta + s_\theta$ $-\rho_0^{-1}(\rho_0 \langle w^* q^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' q' \rangle^p)_z$ $p_z = \theta$ $u_X + \rho_0^{-1}(\rho_0 w)_z = 0$ $Dq/DT + uQ_X + wQ_z - Q_0 w = -s_\theta$ $-\rho_0^{-1}(\rho_0 \langle w^* q^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' q' \rangle^p)_z$	$u, V, p, \theta, q \text{ from } \mathcal{O}(\varepsilon)$ $w, s_\theta \text{ from } \mathcal{O}(\varepsilon^2)$
Synoptic fluctuations in space or time	$u_t^* + Uv_x^* + w^*U_z = -p_x^* - \rho_0^{-1}(\rho_0 \langle w' u' \rangle^s)_z$ $v_t^* + Uv_x^* = -\rho_0^{-1}(\rho_0 \langle w' v' \rangle^s)_z$ $\theta_t^* + U\theta_x^* + w^*\Theta_z + N^2 w^* = -\rho_0^{-1}(\rho_0 \langle w' \theta' \rangle^s)_z + s_\theta^*$ $p_z^* = \theta^*$ $u_x^* + \rho_0^{-1}(\rho_0 w^*)_z = 0$ $q_t^* + Uq_x^* + w^*Q_z - Q_0 w^* = -s_\theta^* - \rho_0^{-1}(\rho_0 \langle w' q' \rangle^s)_z$	All variables from $\mathcal{O}(\varepsilon)$
Mesoscale fluctuations in space and time	$u'_\tau + Uu'_{x'} + w'U_z = -p'_{x'}$ $v'_\tau + Uv'_{x'} = 0$ $\theta'_\tau + U\theta'_{x'} + w'\Theta_z + N^2 w' = s'_\theta$ $p'_z = \theta'$ $u'_{x'} + \rho_0^{-1}(\rho_0 w')_z = 0$ $q'_\tau + Uq'_{x'} + w'Q_z - Q_0 w' = -s'_\theta$	$u', v', p', \theta', q' \text{ from } \mathcal{O}(\varepsilon)$ $w', s'_\theta \text{ from } \mathcal{O}(1)$

TABLE 3. Multiscale asymptotic model across planetary, synoptic, and mesoscales in the order-one rotation regime [regime 2; $\hat{f}$ from $\mathcal{O}(1)$].

Space and time scales	Governing equations	Variables
Trade winds (planetary/intraseasonal)	$DU/DT - \hat{f}V = -P_X - \hat{d}(U - U_0)$ $\hat{f}U = -\varepsilon^2 \hat{d}V$ $D\Theta/DT + N^2W = -\hat{d}_\theta \Theta + S_\theta$ $P_z = \Theta$ $U_X + \rho_0^{-1}(\rho_0 W)_z = 0$ $DQ/DT - Q_0 = -S_\theta$	$\hat{d}, \hat{d}_\theta$ from $\mathcal{O}(\varepsilon)$ $D/DT = \partial/\partial T + U(\partial/\partial X) + W(\partial/\partial Z)$ U, P, Θ, Q from $\mathcal{O}(1)$ V, W, S_θ from $\mathcal{O}(\varepsilon)$ $N^2 = 1$
Planetary/intraseasonal anomalies from the climatology	$Du/DT + uU_X + wU_z - \hat{f}v = -p_X - \hat{d}u$ $-\rho_0^{-1}(\rho_0 \langle w^* u^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' u' \rangle^p)_z$ $\hat{f}u = -\varepsilon^2 \hat{d}v$ $D\theta/DT + u\Theta_X + w\Theta_z + N^2w = -\hat{d}_\theta \theta + s_\theta$ $-\rho_0^{-1}(\rho_0 \langle w^* \theta^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' \theta' \rangle^p)_z$ $p_z = \theta$ $u_X + \rho_0^{-1}(\rho_0 w)_z = 0$ $Dq/DT + uQ_X + wQ_z - Q_0w = -s_\theta$ $-\rho_0^{-1}(\rho_0 \langle w^* q^* \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle w' q' \rangle^p)_z$	u, p, θ, q from $\mathcal{O}(\varepsilon)$ v, w, s_θ from $\mathcal{O}(\varepsilon^2)$
Synoptic fluctuations in space or time	$u_i^* + Uu_x^* + w^*U_z - \hat{f}v^* = -p_x^* - \rho_0^{-1}(\rho_0 \langle w' u' \rangle^s)_z$ $v_i^* + Uv_x^* + \hat{f}u^* = -\rho_0^{-1}(\rho_0 \langle w' v' \rangle^s)_z$ $\theta_i^* + U\theta_x^* + w^*\Theta_z + N^2w^* = -\rho_0^{-1}(\rho_0 \langle w' \theta' \rangle^s)_z + s_\theta^*$ $p_z^* = \theta^*$ $u_z^* + \rho_0^{-1}(\rho_0 w^*)_z = 0$ $q_i^* + Uq_x^* + w^*Q_z - Q_0w^* = -s_\theta^* - \rho_0^{-1}(\rho_0 \langle w' q' \rangle^s)_z$	All variables from $\mathcal{O}(\varepsilon)$
Mesoscale fluctuations in space and time	$u'_\tau + Uu'_{x'} + w'U_z = -p'_{x'}$ $v'_\tau + Uv'_{x'} = 0$ $\theta'_\tau + U\theta'_{x'} + w'\Theta_z + N^2w' = s'_\theta$ $p'_z = \theta'$ $u'_{x'} + \rho_0^{-1}(\rho_0 w')_z = 0$ $q'_\tau + Uq'_{x'} + w'Q_z - Q_0w' = -s'_\theta$	u', v', p', θ', q' from $\mathcal{O}(\varepsilon)$ w', s'_θ from $\mathcal{O}(1)$

In reality, synoptic-scale variability of tropical convection is mainly contributed by CCEWs (Kiladis et al. 2009). The corresponding geophysical flows on the β plane have a velocity scale of dry Kelvin waves $c = NH/\pi$ and a characteristic length scale of Rossby

deformation radius $L_E = (NH/\pi\beta)^{1/2}$, where N is Brunt–Väisälä frequency, H is troposphere height, and β is the Rossby parameter. Therefore, the synoptic time scale in the tropics, $T_E = L_E/c = (c\beta)^{-1/2}$, is about 8 h (Majda 2007), which is equivalent to the

TABLE 4. Multiscale asymptotic model across planetary, synoptic, and mesoscales in the strong rotation regime [regime 3; $\hat{f}$ from $\mathcal{O}(\varepsilon^{-1})$].

Space and time scales	Governing equations	Variables
Planetary/synoptic-time circulation	$U_t - \hat{f}V = -P_X - \hat{d}(U - U_0)$ $\hat{f}U = -\varepsilon^2 \hat{d}V$ $\Theta_t + N^2W = S_\theta - \hat{d}_\theta \Theta$ $P_z = \Theta$ $U_X + \rho_0^{-1}(\rho_0 W)_z = 0$	$\hat{d}, \hat{d}_\theta$ from $\mathcal{O}(1)$ U, P, Θ, S_θ from $\mathcal{O}(1)$ V, W from $\mathcal{O}(\varepsilon)$
Synoptic fluctuations in QG regime	$(D/Dt)(\phi_{xx}^* + \hat{f}^2 \phi_{zz}^*) = -\hat{d}\phi_{xx}^* - \hat{d}_\theta \hat{f}^2 \phi_{zz}^* + \hat{f}(s_\theta^*)_z$	$D/Dt = \partial/\partial t + U(\partial/\partial x)$ $u^* = 0, v^* = \phi_x^*, \theta^* = \hat{f}\phi_z^*$ All variables from $\mathcal{O}(1)$
Mesoscale fluctuations in space and time	$u'_\tau + (U + u^*)u'_{x'} + w'(U + u^*)_z - \hat{f}v' = -(p')_{x'}$ $v'_\tau + (U + u^*)v'_{x'} + w'v'_z + \hat{f}u' = 0$ $N^2w' = s'_\theta$ $u'_{x'} + \rho_0^{-1}(\rho_0 w')_z = 0$	u', v' from $\mathcal{O}(\varepsilon)$ w' from $\mathcal{O}(1)$ p' from $\mathcal{O}(\varepsilon^2)$ s'_θ from $\mathcal{O}(\varepsilon^{-1})$

FIG. 1. Hovmöller diagrams of brightness temperature (K) in cases with various magnitudes of rotation, corresponding to the cases with f at latitude (a) 0° , (b) 1° , (c) 3° , (d) 5° , (e) 9° , (f) 14° , (g) 20° , and (h) 27°N . Depending on the magnitude of rotation, (a)–(d), (e)–(g), and (h) belong to the weak, order-one, and strong rotation regime, respectively. The output is coarse-grained into 16-km grid resolutions (averaged over every eight 2-km longitude grids).

reciprocal of Coriolis frequency f at the latitude 14°N . It describes the characteristic time scale during which synoptic-scale oscillation patterns vary in a magnitude of order one. As shown by Table 1, we divide all rotating scenarios into three regimes, including 1) the ITCZ regime with weak rotation (0° – 6°N), 2) the Indian monsoon trough regime with order-one rotation (6° – 20°N), and 3) the midlatitude regime with strong rotation ($>20^\circ\text{N}$). We choose these three rotation regimes not only because of the observation that convection exhibits distinct characteristics in the tropics, subtropics, and midlatitudes, but also the

different properties of governing equations as shown in Tables 2–4. Besides, the second regime is referred to as the order-one rotation regime, because the Coriolis frequency in this regime corresponds to a time scale at the same order as the typical synoptic time scale (8 h) as mentioned above.

3. Effects of rotation on the multiscale organization of convection

In this section, we first study the spatiotemporal variability of brightness temperature and 850-hPa

FIG. 1. (Continued)

zonal winds, which represent thermodynamic and dynamic aspects of convection, respectively. Notably, Figs. 1 and 2 show that planetary-scale organization of convection only arises in the weak rotation regime, while synoptic-scale organization persists in the order-one rotation regime but diminishes in the strong rotation regime. The effects of rotation on zonal-mean climatology of flow fields are also investigated.

a. Spatiotemporal variability of brightness temperature and 850-hPa zonal winds

Figure 1a shows the Hovmöller diagram of brightness temperature in the nonrotating case, which is the same as Brenowitz et al. (2018). In the first

5 days, numerous westward-moving MCSs are organized into a few eastward-moving synoptic-scale envelopes. After that, a planetary-scale envelope of convection at wavenumber 2 gradually forms and propagates eastward at a speed of 7 m s^{-1} . The 7 m s^{-1} propagation speed corresponds to 17 m s^{-1} relative to the background wind (i.e., -10 m s^{-1} easterly flow), comparable to the speed of convectively coupled gravity waves with the second-baroclinic heating (Takayabu 1994; Wu et al. 2000; Kiladis et al. 2009). Figures 1b–h are for the remaining seven cases (last two cases in the strong rotation regime are not shown). In the weak rotation regime, the planetary-scale organization of convection arises at 1°N in Fig. 1b but diminishes in Figs. 1c and 1d. In contrast, Figs. 1e–g show that synoptic-scale

FIG. 2. Log-scale wavenumber–frequency spectra of (left) brightness temperature (K) and (right) 850-hPa zonal velocity (m s^{-1}) in cases with various magnitude of rotation based on the last 80-day output, corresponding to the cases with f at latitude (a),(b) 0° , (c),(d) 1° , (e),(f) 3° , (g),(h) 5° , (i),(j) 9° , (k),(l) 14° , (m),(n) 20° , (o),(p) 27° , and (q),(r) 35°N . The value at the origin (zonal and time mean) is removed. The horizontal line in each panel indicates the corresponding inertial period $2\pi/f$.

envelopes with embedded westward-moving MCSs dominate in the order-one rotation regime, resembling the two-scale organization of convection in Grabowski and Moncrieff (2001). As the magnitude of rotation increases, the length scale of synoptic-scale envelopes becomes smaller, while their propagation speed is faster. At 27°N in Fig. 1h in the strong rotation regime, scattered MCSs prevail over the whole domain, which is akin to the midlatitude case in Liu and Moncrieff (2004). The typical speed of the simulated synoptic-scale convective organizations in Figs. 1e–h is between 10 and 15 m s^{-1} , while that of MCSs is about 10 m s^{-1} , close to the background easterly flow

(see Fig. 4e). The increasing magnitude of rotation speeds up the propagation of synoptic-scale envelopes but not MCSs. This is consistent with the fact that the Coriolis force mostly affects large-scale circulation (Majda 2007).

Figure 2a shows the wavenumber–frequency spectra of brightness temperature in the nonrotating case. The spectra of brightness temperature is dominated by a peak at wavenumber 2 and a period of 26.7 days, which further extends to larger wavenumbers and shorter periods along a straight line across the origin. Figure 2b shows the spectra of 850-hPa zonal velocity, which is similar to Fig. 2a but with the significant

FIG. 2. (Continued)

spectra of westward-moving modes at wavenumber 1–5. Figures 2c–r are for the remaining eight cases with the rotation. The spectra diagrams for the 45°N case are similar to Figs. 2q and 2r and thus are not shown here. As the magnitude of rotation increases in the weak rotation regime, the spectra accounting for eastward-moving envelopes gradually shift to smaller spatial and temporal scales in Figs. 2c–j. It is worth mentioning that the period of eastward-moving envelopes is close to the corresponding inertial period $2\pi/f$ in the weak rotation regime in Figs. 2c–j, but longer than the inertial period in the order-one and strong rotation regimes in Figs. 2k–r. Overall, the maximum strength of spectra decays gradually as the magnitude of rotation increases, indicating the diminishing spatiotemporal variability of convection. Besides, the spectra band of westward-moving

modes shifts along with the peak of eastward-moving envelopes, reflecting the modulation effect by the latter.

b. Zonal-mean climatology of winds, moisture, and (equivalent) potential temperature

Figure 3 shows the zonal-mean climatology of zonal and meridional velocity, density, water vapor, and (equivalent) potential temperature. As the magnitude of rotation increases, the vertical shear of zonal winds in Fig. 3a diminishes gradually, while the magnitude of meridional winds in Fig. 3b strengthens. The Coriolis force tends to transfer zonal momentum to meridional momentum, resulting in weaker vertical shear of zonal winds that are deviated from the background easterly winds. Vertical profiles of density, potential temperature are mostly similar among all cases in

FIG. 2. (Continued)

Figs. 3c and 3d. As the magnitude of rotation increases, the lower and middle troposphere become moister near 700 hPa in Fig. 3e, resulting in larger midlevel equivalent potential temperature in Fig. 3f. In terms of convective instability, such a thermodynamic mean sounding in Figs. 3d–f provides favorable conditions for deep convection based on the following considerations. For undilute ascent, air parcels lifted up from the surface will obtain larger convective available potential energy (CAPE) in larger rotation cases, solely due to more moisture near the surface as shown in Fig. 3e. For dilute ascent, the entrainment of more moist midlevel environmental air in the larger rotation cases will further increase the parcel buoyancy and promote deeper convection. That said, it is worth noting that the thermodynamic background

soundings among these 10 cases only differ by a small amount so their overall impact on convection may be similar.

c. Multiscale decomposition of flow fields across planetary, synoptic, and mesoscales

To facilitate diagnostic analysis for multiscale interactions in the following sections, we introduce a multiscale decomposition method based on the coarse-graining technique, a straightforward generalization of asymptotic averaging operators (Majda 2007) in a finite domain with small grid spacing. The detailed procedure for decomposing total fields into domain mean, and planetary-, synoptic-, mesoscale fluctuations is explained below. Suppose ϕ is the total field and ϕ_{res} is the residual. Initially, let $\phi_{\text{res}} = \phi$.

FIG. 3. Domain-mean climatology of (a) zonal velocity, (b) meridional velocity, (c) air density, (d) potential temperature, (e) water vapor, and (f) equivalent potential temperature in 10 cases based on the last 80-day output. The horizontal axis shows the value of each field with its dimensional unit given in the panel title.

Step 1: Calculate the mean value of ϕ_{res} in the whole domain and denote it as $\bar{\phi}$ for the domain mean.

Step 2: Update the residual, $\phi_{\text{res}} = \phi - \bar{\phi}$, calculate the mean value of ϕ_{res} over a coarse grid with 2000-km spacing, and denote it as ϕ^p for planetary-scale fluctuations.

Step 3: Update the residual, $\phi_{\text{res}} = \phi - \bar{\phi} - \phi^p$, calculate the mean value of ϕ_{res} over a coarse grid with 256-km spacing, and denote it as ϕ^* for synoptic-scale fluctuations.

Step 4: Update the residual, $\phi_{\text{res}} = \phi - \bar{\phi} - \phi^p - \phi^*$, calculate the mean value of ϕ_{res} over a coarse grid with 16-km spacing, and denote it as ϕ' for meso-scale fluctuations.

The coarse grid spacing (2000, 256, 16 km) is chosen so that 10 coarse grids (20 000, 2560, 160 km) are able to resolve planetary-, synoptic-, and mesoscale fluctuations, respectively. In practice, we first coarse grain the total fields onto coarse grids of 16 km to save

computing expense and filter out fluctuations on smaller scales below 16 km. Such a residual-based technique for multiscale decomposition is similar to that in Brenowitz et al. (2018), except that the latter uses the low-pass filter in the Fourier domain. Figure 4 gives an example for decomposing brightness temperature from the nonrotating case by using this multiscale decomposition method. This method successfully captures the spatiotemporal variability of convection across multiple scales, including eastward-moving planetary-scale envelopes in Fig. 4b, synoptic-scale eastward- and westward-moving disturbances in Fig. 4d and prevalent westward-moving MCSs in Fig. 4e. The domain-mean field in Fig. 4c is steady with negligible variance.

4. The ITCZ regime with weak rotation

In this section, we focus on the ITCZ regime with weak rotation (0° – 6°N). Typical regions in this regime

FIG. 4. Multiscale decomposition of brightness temperature field (K) in the nonrotating case through coarse-graining method for the (a) total field, (b) planetary fluctuations, (c) domain-mean, (d) synoptic fluctuations, and (e) mesoscale fluctuations. Coarse grid size in these panels is (a) 16, (b) 2048, (d) 256, and (e) 16 km.

include the warm pool region from the Indian Ocean to the west Pacific and the ITCZ region over the east Pacific (Waliser and Gautier 1993; Yang et al. 2017). Here we first derive a multiscale model with weak rotation across the planetary-, synoptic-, and mesoscales by following the systematic multiscale asymptotic theory (Majda and Klein 2003; Majda 2007). Then we use it as a diagnostic framework for energy budget analysis to understand why planetary-scale organization of convection diminishes in this regime, as shown by Figs. 1a–d.

a. A multiscale model with weak rotation for interactions of convection across planetary, synoptic, and mesoscales

In general, multiscale asymptotic models are useful for capturing leading-order scale interactions of

convection across multiple spatial and temporal scales (Yang and Majda 2014; Majda and Yang 2016; Yang et al. 2017). The derivation of this multiscale model starts from the 2D anelastic primitive equations on the f plane. The Froude number $\varepsilon = 0.1$ is chosen as the small parameter for multiscale asymptotic analysis. According to the standard scaling (Majda 2007), synoptic-scale spatial and temporal coordinates (x, t) have dimensional units of (1500 km, 8.3 h). Correspondingly, the planetary-scale spatial and temporal coordinates (X, T) have dimensional units (15000 km, 3 days) that are $1/\varepsilon = 10$ times as large as synoptic scales, while mesoscale coordinates (x', τ) are $\varepsilon = 1/10$ of synoptic scales. As for physical variables, zonal and meridional velocity, (u, v) , are scaled in a unit of 50 m s^{-1} , and vertical

velocity w in a unit of 0.16 m s^{-1} . Pressure perturbation p is scaled in a unit of $2500 \text{ m}^{-2} \text{ s}^{-2}$, potential temperature anomalies θ and moisture anomalies q in a unit of 15 K , and diabatic heating s_θ in a unit of 45 K day^{-1} . The order of variables are summarized in the third column of Table 2. To separate terms into different scales, spatial averaging operator $\bar{u}$ and temporal averaging operator $\langle u \rangle$ for an arbitrary variable u , and the superscripts p and s indicate the averaging on planetary and synoptic scales, respectively.

This multiscale model consists of four groups of equations, each of which governs dynamics on one specific spatial temporal scales. In detail, the first group of equations in the third row of Table 2 describe trade wind dynamics on the planetary/intraseasonal scale as a climatological background. In contrast, the second group of equations in the fourth row describes the planetary/intraseasonal anomalies under the effects of rotation, which are also influenced by the advection of background flow U , W and interaction terms involving trade wind fields U , Θ , Q . Furthermore, the eddy transfer of zonal momentum from synoptic fluctuations $-\rho_0^{-1}(\rho_0 \langle \overline{w^* u^*} \rangle^p)_z$ and that from mesoscale fluctuations $-\rho_0^{-1}(\rho_0 \langle \overline{w' u'} \rangle^p)_z$ represent upscale impact of synoptic- and mesoscale dynamics. Similar eddy terms also appear at the right-hand side of meridional momentum, potential temperature, and moisture equations. The third group of equations in the fifth row governs the dynamics of synoptic-scale fluctuations, which is affected by the trade wind fields as well as eddy terms from mesoscale fluctuations. The last group of equations in the sixth row describes the dynamics of mesoscale fluctuations advected by trade wind fields.

b. Effects of eddy momentum transfer on planetary-scale momentum and kinetic energy budget

According to the governing equations for planetary-scale zonal and meridional momentum in Table 2,

$$\frac{Du}{DT} + uU_x + wU_z - \hat{f}V = -p_x - \hat{d}u - \rho_0^{-1}(\rho_0 \langle \overline{w^* u^*} \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle \overline{w' u'} \rangle^p)_z, \quad (1)$$

$$\frac{DV}{DT} + \hat{f}u = -\hat{d}V - \rho_0^{-1}(\rho_0 \langle \overline{w^* v^*} \rangle^p)_z - \rho_0^{-1}(\rho_0 \langle \overline{w' v'} \rangle^p)_z, \quad (2)$$

where the trade wind background U is assumed to be -10 m s^{-1} . After taking the climatological-mean $[\cdot]$ (zonal and time averaging), the above equations are rewritten as

$$[u_T] = [\hat{f}V] + [-\hat{d}u] + [-\rho_0^{-1}(\rho_0 \langle \overline{w^* u^*} \rangle^p)_z] + [-\rho_0^{-1}(\rho_0 \langle \overline{w' u'} \rangle^p)_z], \quad (3)$$

$$[V_T] = [-\hat{f}u] + [-\hat{d}V] + [-\rho_0^{-1}(\rho_0 \langle \overline{w^* v^*} \rangle^p)_z] + [-\rho_0^{-1}(\rho_0 \langle \overline{w' v'} \rangle^p)_z], \quad (4)$$

which indicate that eddy momentum transfer from synoptic- and mesoscale fluctuations influences the planetary-scale winds.

Figures 5a–c show the climatological-mean vertical profiles of eddy zonal momentum transfer from meso-, synoptic-, and planetary-scale fluctuations. In detail, the eddy momentum transfer from mesoscale fluctuations in Fig. 5a induces westward (eastward) momentum forcing in the lower (middle and upper) troposphere. This vertical distribution of momentum transport tendency for westward-propagating MCSs is consistent with the Moncrieff's archetype model (Moncrieff 1981, 1992; Moncrieff et al. 2017). [See Fig. 12 in Moncrieff et al. (2017), but note that the schematic is for eastward-propagating MCSs.] In contrast, eddy momentum transfer from synoptic-scale fluctuations in Fig. 5b is negligible, while that from planetary-scale fluctuations in Fig. 5c has significant momentum forcing only above 600 hPa. In addition, Figs. 5d and 5e show the Coriolis force term and momentum drag, both of which have the opposite vertical profiles as that in Fig. 5a. As the latitude increases, the momentum damping effect in Fig. 5d gets strengthened, while that in Fig. 5e gets weakened. Figures 6a–c shows the climatological-mean vertical profiles of eddy meridional momentum transfer from meso-, synoptic- and planetary-scale fluctuations. In particular, the eddy meridional momentum transfer from mesoscale fluctuations induces both low-level and midtropospheric southward momentum forcing and upper-tropospheric northward momentum forcing, while that from synoptic fluctuations is negligible. The Coriolis force and momentum damping in Figs. 6d and 6e have the similar vertical profiles but in the opposite sign.

After multiplying Eqs. (1) and (2) by $\rho_0 u$ and $\rho_0 v$, respectively, and taking climatological mean, we can obtain the planetary kinetic energy budget equations,

$$\left[\left(\frac{1}{2} \rho_0 u^2 \right)_T \right] = [\rho_0 \hat{f}Vu] + [-\rho_0 p_x u] + [-\hat{d} \rho_0 u^2] + [-(\rho_0 \langle \overline{w^* u^*} \rangle^p)_z u] + [-(\rho_0 \langle \overline{w' u'} \rangle^p)_z u], \quad (5)$$

FIG. 5. Vertical profiles of climatological-mean (domain-mean and time-mean) zonal momentum budget terms (m s^{-2}) based on the last 80-day model output in the weak rotation regime. (top) Eddy zonal momentum transfer from (a) mesoscale fluctuations, (b) synoptic fluctuations, and (c) planetary fluctuations. (f) The total eddy zonal momentum transfer [i.e., (a) + (b) + (c)]. Also shown are (d) The Coriolis term and (e) momentum damping.

$$\left[\left(\frac{1}{2} \rho_0 V^2 \right)_T \right] = [-\rho_0 \hat{f} u V] + [-\hat{d} \rho_0 V^2] \\ + [-(\rho_0 \langle \overline{w^* v^*} \rangle_z)^P V] + [-(\rho_0 \langle \overline{w' v'} \rangle_z)^P V]. \quad (6)$$

Figures 7a–c show the vertical profiles of energy source and sink terms in the planetary-scale kinetic energy budget for zonal winds. Figure 7a shows the deceleration term involving the Coriolis force, which transfers kinetic energy from zonal winds to meridional winds. In contrast, both terms involving synoptic- and mesoscale fluctuations in Figs. 7b and 7c induce acceleration effects in both lower and upper troposphere,

whose magnitudes decrease gradually as the latitude increases. Figures 7d–f are for meridional winds. The term involving eddy meridional momentum transfer from mesoscale fluctuations in Fig. 7f always induces deceleration effects throughout the troposphere.

Figures 8a and 8b show the planetary-scale kinetic energy budget for zonal and meridional winds. As the latitude increases from 0° to 1° , 3° , 5° in Fig. 8a, acceleration effect induced by both eddy momentum transfer terms gets weakened. In contrast, the Coriolis force term increases dramatically from zero at 0° latitude to nonzero at 1°N , but maintains a fairly constant value as the latitude further increases at 3° and 5°N . Besides, both the terms involving pressure gradient and damping

FIG. 6. Vertical profiles of climatological-mean (domain-mean and time-mean) meridional momentum budget terms (m s^{-2}) based on the last 80-day model output in the weak rotation regime. (top) Eddy meridional momentum transfer from (a) mesoscale fluctuations, (b) synoptic fluctuations, and (c) planetary fluctuations. (f) The total eddy meridional momentum transfer shows total [i.e., (a) + (b) + (c)]. Also shown are (d) the Coriolis term and (e) momentum damping.

decrease as the latitude increases. As shown by Fig. 8b, the term involving eddy meridional momentum transfer from synoptic-scale fluctuations induces weak acceleration effect, while that from mesoscale fluctuations and the damping term induce significant deceleration effect. It is worth mentioning that this energy budget analysis is based on solutions in the equilibrium state so that all terms are balanced with each other with negligible time tendency. In literature, the transient behavior of solutions was discussed by using budget analysis of moist static energy variance (Wing and Emanuel 2014). To explain how the convective organization develops in the first place, it is better to refer to the solutions at the early stage of the simulations. As shown by Fig. 1a, numerous westward-moving mesoscale

disturbances are first generated during the first 2 days in the presence of background easterly winds and warm SST surface. Between day 3 and day 10, these mesoscale disturbances are embedded in a few eastward-moving synoptic-scale convective envelopes, resembling the simulations of Grabowski and Moncrieff (2001). The time series of kinetic energy budget terms at the early stage indicates that eddy momentum transfer of mesoscale fluctuations induces acceleration effects to planetary zonal winds at the first 5 days (see Fig. S1 in the online supplemental material). Similar to Fig. 14, it is the eddy momentum transfer from mesoscale fluctuations that accelerates synoptic-scale zonal winds and promotes synoptic-scale organization of convection at the very beginning. Then in the presence of meso- and

FIG. 7. Vertical profiles of climatological-mean (domain-mean and time-mean) planetary-scale kinetic energy source and sink terms ($\text{kg m}^{-1} \text{s}^{-3}$), based on the last 80-day model output in the weak rotation regime. (top) Profiles for (a) the Coriolis force, (b) eddy zonal momentum transfer from synoptic fluctuations, and (c) eddy zonal momentum transfer from mesoscale fluctuations. (d)–(f) As in (a)–(c), but for meridional winds.

synoptic-scale organization of convection, planetary-scale organization of convection develops gradually afterward, due to the acceleration effects from eddy momentum transfer from both meso- and synoptic-scale fluctuations. Last, the negative budget residual in Figs. 8a and 8b includes a linear momentum damping term, representing zonal and meridional momentum dissipation (i.e., cumulus drag) in the tropical atmosphere.

Figure 9a shows the schematic diagram for planetary-scale kinetic energy budget in the weak rotation regime. According to Fig. 8a, the dominant acceleration effect comes from the term involving eddy zonal momentum transfer from mesoscale fluctuations $[-(\rho_0 \langle w'u' \rangle^P)_z u]$, while the dominant deceleration effect comes from the

term involving the Coriolis force $[\rho_0 \hat{f} V u]$. As the magnitude of rotation increases, this acceleration effect decreases dramatically while the deceleration effect increases instead. The resulting reduced planetary-scale kinetic energy budget of zonal winds explains the diminishing planetary-scale organized convection.

Both changed acceleration/deceleration effects should be traced back to the increasing magnitude of rotation, as it is the only difference in the model input. In fact, the increasing deceleration term $[\rho_0 \hat{f} V u]$ can be simply explained by the larger value of f at higher latitudes. As for the acceleration term $[-(\rho_0 \langle w'u' \rangle^P)_z u]$, its decreasing strength is attributed to less favorable conditions for MCSs provided by weaker background vertical shear of

FIG. 8. Climatological-mean (zonal- and vertical-mean, and time-mean) total planetary-scale kinetic energy source and sink terms ($\text{kg m}^{-1} \text{ s}^{-3}$) for (a) zonal winds, (b) meridional winds, based on the last 80-day model output in the weak rotation regime. The y-axis limits in both panels are $2.35 \times 10^{-5} \text{ kg m}^{-1} \text{ s}^{-3}$.

zonal winds as shown in Fig. 9b. According to Fig. 5d, the Coriolis term fV induces a momentum forcing in the opposite sign as the climatological-mean zonal winds in Fig. 3a, resulting in reduced low-level vertical shear.

c. *Effects of eddy heat transfer on planetary-scale heat and available potential energy budget*

The governing equation for planetary-scale potential temperature anomalies in Table 2 reads as follows:

$$\theta_T + U\theta_X + N^2 w = -\hat{d}_\theta \theta - \rho_0^{-1} (\rho_0 \langle w^* \theta^* \rangle^p)_z - \rho_0^{-1} (\rho_0 \langle w' \theta' \rangle^p)_z + s_\theta, \quad (7)$$

where the trade wind background is assumed to be $U = -10 \text{ m s}^{-1}$ and $\Theta = 0 \text{ K}$. The corresponding climatological-mean equation is,

$$[\theta_T] = [-N^2 w] + [-\hat{d}_\theta \theta] + [-\rho_0^{-1} (\rho_0 \langle w^* \theta^* \rangle^p)_z] + [-\rho_0^{-1} (\rho_0 \langle w' \theta' \rangle^p)_z] + [s_\theta]. \quad (8)$$

Figure 10 show the climatological-mean vertical profiles of eddy heat transfer from meso-, synoptic-, and planetary-scale fluctuations. Unlike Figs. 5 and 6, the

vertical profiles of all eddy terms do not change much as the latitude increases, indicating that these terms are not directly responsible for the diminishment of planetary-scale organization of convection. In fact, both eddy heat transfer from synoptic- and mesoscale fluctuations introduce heating in the lower troposphere and increase CIN, providing unfavorable conditions for convection.

After multiplying Eq. (7) by $\rho_0 \theta / N^2$ and taking climatological mean, the governing equation for available potential energy budget is obtained below as

$$\left[\left(\rho_0 \frac{\theta^2}{2N^2} \right)_T \right] = [-\rho_0 w \theta] + \left[-\rho_0 \hat{d}_\theta \frac{\theta^2}{N^2} \right] + \left[-(\rho_0 \langle w^* \theta^* \rangle^p)_z \frac{\theta}{N^2} \right] + \left[-(\rho_0 \langle w' \theta' \rangle^p)_z \frac{\theta}{N^2} \right] + \left[\rho_0 s_\theta \frac{\theta}{N^2} \right], \quad (9)$$

where the term $[-\rho_0 w \theta]$ transfers energy between kinetic energy and available potential energy.

Figure 11 shows the climatological-mean vertical profiles of energy source and sink terms in available

(a) Planetary kinetic energy of zonal winds diminishes as rotation f increases (b) Mesoscale convective systems diminishes as rotation f increases

FIG. 9. Schematic diagram explaining why planetary-scale kinetic energy of zonal winds diminishes as the rotation f increases in the weak rotation regime. (a) Acceleration/deceleration effects in the planetary-scale kinetic energy budget of both zonal and meridional winds, where thick (thin) arrows indicate the dominant (secondary) energy source/sink terms. The red up (blue down) arrow represents increasing (decreasing) magnitude. Overall, the diminishment of planetary kinetic energy of zonal winds is due to 1) an increasing deceleration term involving the Coriolis force and 2) a decreasing acceleration term involving eddy zonal momentum transfer from mesoscale fluctuations. (b) Diminishment of mesoscale convective systems to the decreasing low-level vertical shear in the background sounding as the rotation f increases.

potential energy budget. As shown by Figs. 11b and 11c, the energy source/sink terms involving eddy heat transfer from synoptic- and mesoscale fluctuations share the similar vertical profiles, both of which feature an energy source below 850 hPa and above 300 hPa, and an energy sink between 350 and 850 hPa. Meanwhile, neither term changes much throughout the troposphere as the latitude increases, indicating that these terms are not directly responsible for the diminishing planetary-scale organization.

d. Effects of eddy transfer of equivalent potential temperature on the planetary-scale atmospheric stability

Similar to Eq. (8), the governing equation for equivalent potential temperature θ_e reads as follows:

$$[(\theta_e)_T] = [-N_e^2 w] + [-\hat{\partial}_\theta \theta_e] + [-\rho_0^{-1} (\rho_0 \overline{w^* \theta_e^*})^p] + [-\rho_0^{-1} (\rho_0 \overline{w' \theta_e'})^p], \quad (10)$$

where N_e represents background stratification of equivalent potential temperature.

Figure 12 shows the climatological-mean vertical profiles of eddy transfer of equivalent potential temperature from planetary-, synoptic-, and mesoscale fluctuations. Among these three terms, eddy terms from synoptic- and mesoscale fluctuations dominate and induce cooling and drying effects below 850 hPa and heating and moistening effects above that level. The eddy term from planetary fluctuations have negligible magnitude throughout the troposphere. It is worth mentioning that these vertical

FIG. 10. Vertical profiles of climatological-mean (domain-mean and time-mean) eddy heat transfer (K s^{-1}) from (a) mesoscale fluctuations, (b) synoptic fluctuations, (c) planetary fluctuations, and (d) the total, based on the last 80-day model output in the weak rotation regime.

profiles do not change as the magnitude of rotation increases.

5. The Indian monsoon regime with order-one rotation

In this section, we will focus on the Indian monsoon regime with order-one rotation (6° – 20°N). A typical region in this regime is the monsoon trough over the Indian subcontinent (Gadgil 2003; Yang et al. 2019a). As shown by Figs. 1e–g, large-scale convection is

dominated by synoptic-scale envelopes that move eastward at a speed of 15 m s^{-1} , resembling the simulation by Grabowski and Moncrieff (2001). It is important to investigate the upscale impact of MCSs on synoptic-scale dynamics and understand why synoptic-scale organization persists in this regime.

Table 3 shows the multiscale model for the scale interactions across meso-, synoptic, and planetary scales in the order-one rotation regime. To derive this multiscale model, we use the same physical scaling for all physical variables as section 4a, except for the Coriolis force parameter f in the

FIG. 11. Vertical profiles of climatological-mean (domain-mean and time-mean) available potential energy source and sink terms ($\text{kg m}^{-1} \text{s}^{-3}$), based on the last 80-day model output in the weak rotation regime. (a) The term involving energy transfer between kinetic energy and available potential energy. (b),(c) Available potential energy source and sinks terms involving eddy heat transfer from (b) synoptic fluctuations and (c) mesoscale fluctuations, respectively. Potential temperature is rescaled by a constant, $\hat{\theta} = (g/\theta)\theta$.

order-one magnitude. Thus, the two models in Tables 2 and 3 share many similar features. The major difference lies in the fact that trade wind background and synoptic-scale dynamics in Table 3 feel the Coriolis force. Moreover, this three-scale model can be regarded as the coupling between the IMMD model (Biello and Majda 2010) for planetary- and synoptic-scale interactions and the MESD model (Majda 2007) for synoptic- and mesoscale interactions.

Upscale impact of mesoscale fluctuations on synoptic-scale dynamics

According to Table 3, synoptic-scale dynamics is driven by eddy transfer of momentum, temperature and moisture from mesoscale fluctuations. It should be interesting to investigate the upscale impact of MCSs on synoptic-scale organization of convection in this regime. The governing equations for synoptic-scale kinetic energy budget of zonal and meridional winds and available potential energy budget read below:

$$\left[\left(\frac{1}{2} \rho_0 (u^*)^2 \right) \right]_t = [\rho_0 \hat{f} v^* u^*] + [-\rho_0 p_x^* u^*] + [-(\rho_0 \langle \overline{w' u'} \rangle^s)_z u^*], \quad (11)$$

$$\left[\left(\frac{1}{2} \rho_0 (v^*)^2 \right) \right]_t = [-\rho_0 \hat{f} v^* u^*] + [-(\rho_0 \langle \overline{w' v'} \rangle^s)_z v^*], \quad (12)$$

$$\left[\left(\rho_0 \frac{(\theta^*)^2}{2N^2} \right) \right]_t = [-\rho_0 w^* \theta^*] + \left[-(\rho_0 \langle \overline{w' \theta'} \rangle^s)_z \frac{\theta^*}{N^2} \right] + \left[\rho_0 s_\theta^* \frac{\theta^*}{N^2} \right]. \quad (13)$$

Figure 13 shows the climatological-mean vertical profiles of energy source/sink terms on the synoptic-scale kinetic and available potential energy budgets. It turns out that eddy zonal momentum transfer in Fig. 13a induces acceleration effects throughout the troposphere, whose magnitude decays gradually as the latitude increases. In contrast, eddy meridional momentum transfer in Fig. 13b induces weaker deceleration effects, while eddy heat transfer in Fig. 13c induces alternate energy source and sink at different levels.

Figure 14 shows the synoptic-scale kinetic energy budget for zonal and meridional winds. The acceleration/deceleration effects induced by the Coriolis force do not change much. As for kinetic

FIG. 12. Vertical profiles of climatological-mean (domain-mean and time-mean) eddy transfer of equivalent potential temperature (K s^{-1}) from (a) mesoscale fluctuations, (b) synoptic fluctuations, (c) planetary fluctuations, and (d) the total, based on the last 80-day model output in the weak rotation regime.

energy of zonal winds in Fig. 14a, the dominant acceleration effect due to eddy zonal momentum transfer from mesoscale fluctuations decays as the latitude increases. Correspondingly, the deceleration effect due to pressure gradient also decays. As for meridional winds, the acceleration effect induced by the Coriolis force is balanced by the term involving eddy meridional momentum transfer and the damping residual. The residual in Fig. 14a has a small but positive value, which may be related to the small horizontal eddy terms that are ignored during the

derivation of the multiscale models. The residual in Fig. 14b is too large to be ignored and behaves as momentum dissipation, presumably due to the frictional effect from unorganized convection that has been excluded in the budget analysis. These residual terms are relatively smaller than the corresponding acceleration terms and thus should not change the leading-order kinetic energy balance.

Figure 15 shows the schematic diagram for synoptic-scale kinetic energy budget. According to Fig. 14a, the dominant acceleration effect in synoptic kinetic

FIG. 13. Vertical profiles of climatological-mean (domain-mean and time-mean) synoptic-scale energy source and sink terms ($10^{-5} \text{ kg m}^{-1} \text{ s}^{-3}$) based on the last 80-day model output in the order-one rotation regime. Panels show the terms involving (a) eddy zonal momentum transfer, (b) eddy meridional momentum transfer, (c) eddy heat transfer, and (d) the Coriolis force, as well as the terms representing (e) energy conversion between kinetic energy and (f) available potential energy.

energy of zonal winds is induced by eddy zonal momentum transfer from mesoscale fluctuations $[-(\rho_0 \langle w' u' \rangle)_z u^*]$, while the deceleration effect comes from the term involving the Coriolis force $[\rho_0 \hat{v} u^*]$. Thus, this acceleration effect maintains the synoptic-scale organization of convection. As the latitude further increases, this acceleration effect decays gradually, while the deceleration effect is unchanged. The resulting reduced synoptic-scale kinetic energy of zonal winds explains the diminishment of synoptic-scale organization in the order-one and strong rotation regimes in Figs. 1e–h. Similar to the weak rotation regime, the decaying upscale impact of MCSs is attributed to less favorable conditions for

convection provided by weaker background vertical shear of zonal winds at higher latitudes, as shown in Fig. 3a.

6. The midlatitude regime with strong rotation

In this section, we consider the midlatitude regime with strong rotation. As shown by Fig. 1h, the solution in this regime is characterized by scattered and random MCSs prevailing in the whole domain. It is interesting to investigate the upscale impact of MCSs and understand the vanishment of synoptic-scale organization of convection in the strong rotation regime.

FIG. 14. Climatological-mean (zonal- and vertical-mean and time-mean) total synoptic-scale kinetic energy source and sink terms ($\text{kg m}^{-1} \text{s}^{-3}$) for (a) zonal winds and (b) meridional winds, based on the last 80-day model output in the order-one rotation regime. The y-axis limit in both panels are $1.1 \times 10^{-5} \text{ kg m}^{-1} \text{s}^{-3}$.

a. A multiscale model with strong rotation for interactions of convection across planetary, synoptic, and mesoscales

It is well known that large-scale circulation at mid-latitudes is governed by the QG dynamics. Thus, the standard QG scaling (Vallis 2017) is adopted here. In details, synoptic-scale spatial and temporal coordinates (x, t) have dimensional units of (1000 km, 28 h). Correspondingly, the planetary-scale spatial coordinate X has dimensional units 10 000 km that are $1/\varepsilon = 10$ times of those on the synoptic scale, while mesoscale coordinates (x', τ) are $\varepsilon = 1/10$ of synoptic-scale ones. As for physical variables, zonal and meridional velocity, (u, v) , are scaled in a unit of 10 m s^{-1} , and vertical velocity w in a unit of 0.1 m s^{-1} . Pressure perturbation p is scaled in a unit of $1000 \text{ m}^{-2} \text{s}^{-2}$, potential temperature anomalies θ and moisture anomalies q in a unit of 3 K, and diabatic heating s_θ in a unit of 2.57 K day^{-1} .

Table 4 shows the multiscale model in this strong rotation regime with three groups of equations, each of which governs one single scale dynamics. In brief, the planetary-scale dynamics is governed by longwave

approximation equations, the synoptic-scale dynamics is governed by QG equations, and the mesoscale dynamics is governed by the linear mesoscale equatorial weak temperature gradient (MEWTG) equations (Majda and Klein 2003; Majda et al. 2008). Notably, this multiscale model is distinguished from the other two models in Tables 2 and 3 by the absence of eddy terms across planetary, synoptic, and mesoscales. This multiscale model predicts theoretically that upscale impact of synoptic- and mesoscale fluctuations is negligible in the strong rotation regime.

b. Upscale impact of mesoscale fluctuations on synoptic-scale dynamics

Figure 16 shows the synoptic-scale kinetic energy budget for zonal and meridional winds in the strong rotation regime. The overall features of all energy source and sink terms are similar to those in Fig. 14. In particular, eddy zonal momentum transfer from mesoscale fluctuations still induces acceleration effect in the kinetic energy budgets, whose magnitude further decreases as the latitude increases. In contrast, eddy meridional momentum transfer induces deceleration effects. However, when

Synoptic kinetic energy of zonal winds diminishes as rotation f increases

FIG. 15. Schematic diagram explaining the maintenance of synoptic organization of convection and its diminishment as the rotation further increases in the order-one rotation regime. This figure shows acceleration/deceleration effects in the synoptic-scale kinetic energy budget of both zonal and meridional winds, where thick (thin) arrows indicate the dominant (secondary) energy source/sink terms. The blue down arrow represents decreasing in magnitude. Overall, the diminishment of synoptic kinetic energy of zonal winds is due to a decreasing acceleration term involving eddy zonal momentum transfer from mesoscale fluctuations. See Fig. 12 for an explanation of the diminishment of mesoscale convective systems.

compared with Fig. 14, these acceleration/deceleration effects are too weak to support synoptic-scale organization of convection. Unlike Fig. 14, the deceleration effect due to the Coriolis force gradually decreases as the rotation increases.

7. Concluding discussion

This study is aimed at investigating the effects of rotation on the multiscale organization of convection with the following goals. First, we use a global 2D CRM to simulate multiscale organization of convection in three rotation regimes (weak, order-one, and strong), representing idealized ITCZ region (0° – 6°N), Indian monsoon region (6° – 20°N), and midlatitude region (20° – 45°N), respectively. Second, we derive a multiscale asymptotic model for upscale and downscale impacts in each rotation regime and use it as a diagnostic framework for energy budget analysis. Third, we explain why planetary-scale organization diminishes in the weak rotation regime as the magnitude of rotation increases and investigate the role of eddy transfer of momentum, temperature, and equivalent potential temperature from meso- and synoptic-scale fluctuations. Last,

we explain why synoptic-scale organization persists in the order-one rotation regime but diminishes in the strong rotation regime.

Here we use the 2D version of the SAM model to simulate multiscale organization of convection with different magnitudes of rotation. In the weak rotation regime, planetary-scale organization of convection arises at latitudes of 0° and 1°N , but diminishes as the magnitude of rotation increases. The eastward-moving planetary-scale envelope contains several eastward- and westward-moving synoptic-scale disturbances with numerous embedded MCSs. In the order-one rotation regime, convection is organized in a two-scale structure with eastward-moving synoptic-scale envelopes and westward-moving embedded MCSs. In the strong rotation regime, numerous scattered and unorganized MCSs prevail in the whole domain. The effect of rotation on large-scale organization of convection as revealed by this CRM simulation is consistent to that in Majda et al. (2015). With both radiative cooling and surface fluxes fixed, the planetary-scale organization of convection in our simulations is mainly due to the multiscale interactions of flow fields, distinguishing itself from

FIG. 16. Climatological-mean (zonal- and vertical-mean and time-mean) total synoptic-scale kinetic energy source and sink terms ($\text{kg m}^{-1} \text{s}^{-3}$) for (a) zonal winds and (b) meridional winds, based on the last 80-day model output in the strong rotation regime. The y-axis limit in both panels is $0.80 \times 10^{-5} \text{ kg m}^{-1} \text{s}^{-3}$.

several previous theories that focus on convection–radiation–surface flux feedbacks (Bretherton et al. 2005; Wing and Emanuel 2014; Bretherton and Khairoutdinov 2015).

Here we divide all scenarios into three regimes (weak, order-one, and strong) in terms of the magnitude of rotation. In each rotation regime, a three-scale model is derived by using the multiscale asymptotic method and used as a diagnostic framework to study the scale interactions of convection across planetary-, synoptic-, and mesoscales. Although they are reduced models from the primitive equations, these multiscale models presumably capture the leading-order quantities of all flow fields with only small errors. The advantages of using these multiscale models as a diagnostic framework for budget analysis lie in three aspects, including 1) modeling the scale interactions of flow fields across multiple scales, 2) highlighting possible dominant terms in the energy budget, and 3) simplifying the diagnostic studies by ignoring secondary terms. By diagnostically calculating energy budget based on these multiscale models, we figure out energy transfer routes on both planetary and synoptic scales and

summarize them in the schematic diagrams in Figs. 9a and 15. As shown by Fig. 9a, planetary kinetic energy of zonal winds is fueled by dominant acceleration effect from MCSs and also that from synoptic convectively coupled waves, but consumed through energy transfer to kinetic energy of meridional winds and available potential energy as well as dissipation. The energy transfer routes on synoptic scale in Fig. 15 are similar to those on planetary scale, reflecting the self-similarity property of convection (Majda 2007). The small but nonnegligible residual reflects a limitation of the current energy budget analysis, which are worth further investigation by adding more relevant forcing and damping terms into the diagnostic framework.

The results here highlight the crucial upscale impact of eddy zonal momentum transfer from mesoscale fluctuations on both planetary- and synoptic-scale organization of convection. As the magnitude of rotation increases, its acceleration effect on the planetary kinetic energy of zonal winds decreases gradually, diminishing the planetary-scale organization of convection. Similarly, due to its decreasing acceleration effect on

synoptic kinetic energy of zonal winds, synoptic-scale organization of convection only persists in the order-one rotation regime but diminishes in the strong rotation regime. This indicates a need to parameterize upscale impact of MCSs in the coarse-resolution GCMs. In fact, the MESD model (Majda 2007) theoretically predicts the significant upscale impact of MCSs on eastward-moving CCKWs (Yang and Majda 2017, 2018) and 2-day waves (Yang and Majda 2019). Based on the explicit expressions of eddy terms obtained from the MESD model, Yang et al. (2019b) proposed a basic parameterization of upscale impact of upshear-moving MCSs and showed that this parameterization significantly improves key features of the MJO analog in a multcloud model. Moncrieff et al. (2017) introduced a parameterization for collective effects of mesoscale organized convection that are missing in the contemporary cumulus parameterization in the GCM.

The diminishing acceleration effects from MCSs are traced back to the increasing magnitude of rotation, since it is the only difference in the model input among all simulations. As the magnitude of rotation increases, vertical shear of zonal winds in the lower troposphere decays, providing less favorable conditions for the generation and propagation of MCSs. Consequently, their upscale impact on the planetary and synoptic kinetic energy diminishes. The schematic diagram in Fig. 9b specifically depicts the effects of increasing rotation on background sounding with less favorable conditions for promoting MCSs. Such upscale and downscale impacts illustrate the crucial role of multiscale interactions in scale selection and organization of convection. Studying the effects of rotation should help improve our fundamental understanding of large-scale organization of convection at different latitudes. Besides, the MCSs in this 2D CRM with rotation share several realistic features with 3D CRMs, while those in 2D CRMs without rotation typically have an unrealistic strong circulation in the zonal direction.

According to Figs. 1 and 3, the rotation modifies both the mean and fluctuations of dynamic and thermodynamic flow fields across multiple spatiotemporal scales. To investigate whether the modified mean flow alone could change the convective organization, we repeat the previous nonrotating case but with reduced momentum nudging time scale (12–24 h). The resulting mean zonal winds get closer to the nudging background easterlies (-10 m s^{-1}), similar to those in the rotating cases as shown in Fig. 3. Despite these mean state changes, these simulations retain a similar degree of organization (not shown). This result suggests that the effect of the rotation on convective perturbations may play a dominant role in influencing the convective organization.

This study can be elaborated and extended in various ways. The implication of multiscale organization of convection presented here is limited due to the 2D model configuration. For example, Held et al. (1993) found model artifacts in their 2D cloud-resolving simulations with a QBO-like oscillation and very strong jets in the troposphere. Liu et al. (2012) reported that the 2D model configuration has limitations in simulating convective organization such as its wave signature and vertical structure. Thus, one research direction is to implement the 3D simulations and investigate the effects of rotation. Meanwhile, the validity of using multiscale asymptotic models as a diagnostic framework depends on appropriate physical scaling for all flow fields and a good multiscale decomposition method for capturing the scale separation property of solutions. Another research direction is to consider the multiscale interactions of convection over the warm pool scenario. Also, it should be interesting to consider the scenario in the presence of active radiation and surface flux and investigate whether the multiscale interaction mechanism would collaborate with the convection–radiation–surface flux feedback mechanisms. Last, the relative importance of low-level thermodynamics and vertical shear could be assessed by performing additional simulations that exclude either of the two. For example, the impact of wind shear is completely excluded if the modeled zonal wind is nudged toward the wind profile in the nonrotation case.

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