MJO Initiation Triggered by Amplification of Upper-Tropospheric Dry Mixed Rossby-Gravity Waves

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Abstract A possibly important dynamical process for the Madden–Julian oscillation (MJO) convective initiation is proposed. An MJO event during the “CINDY2011” field campaign is triggered by eastward-moving lower-tropospheric mixed Rossby-gravity (MRG) wave packets, and its leading precursor is predominance of upper-tropospheric MRGs in the Indian Ocean (IO). Simple three-dimensional model experiments reveal that the upper-tropospheric MRGs in the IO are amplified particularly in the western IO (WIO) by their westward advection and wave accumulation due to the upper-level convergence in mean easterlies of the Walker circulation. The model also predicts downward dispersion of the amplified upper-tropospheric MRGs and resultant lower-tropospheric MRG wave packet formation. This MRG evolution consistently explains the MJO initiation process during CINDY2011, which is further verified by ray tracing for MRGs. Upper-tropospheric circumnavigating Kelvin waves assist the proposed mechanism by promoting MRG-wave accumulation (advection) in their westerly (easterly) phases via enhanced zonal convergence and weakened easterlies (enhanced easterlies).

Plain Language Summary In the tropics, there exists a huge cluster of clouds and rainfall systems moving from the western Indian Ocean (WIO) to the western Pacific, called the Madden–Julian Oscillation (MJO). It is of great interest how and when MJO clouds are formed in the WIO because of their large impacts on global weather patterns, but we have not fully understood it yet. Using a simplified computer simulation and observational data, we find that the formation of MJO tall clouds can start from the “upper-sky” (~10 km altitude) short-period wind variations. During the no-cloud period of MJO, energy of upper-sky wind variations can be input above the central Indian Ocean. Then, it is carried to the WIO and amplified there by upper-sky easterly winds and their convergence associated with the seasonal atmospheric circulation. Because the amplified upper-sky wind energy above the WIO is further dispersed downward, near-surface wind variations become active, which triggers MJO clouds. This new mechanism is theoretically plausible but has been confirmed only in a limited case. It thus should be evaluated for more observations.

1. Introduction
The Madden–Julian oscillation (MJO) is the most prominent intraseasonal variability in the tropics (Madden & Julian, 1972), observed as an eastward-propagating large-scale organized convective system over the Indo-Pacific region. The whole picture of the MJO cannot be explained by classical equatorial wave theories (Matsumo, 1966; Takayabu, 1994; Wheeler & Kiladis, 1999). Also, the MJO has extensive impacts on global weather patterns (Zhang, 2013). It thus has been of great interest to scrutinize the MJO mechanics and to improve the MJO prediction capability over the past decades. In particular, revealing the physics underlying MJO initiation is one of challenging tasks, as inferred from the fact that many pathways to MJO onset in the Indian Ocean (IO) have been proposed (Jiang et al., 2020, and references therein).

MJO initiation processes can be largely divided into two stages; S₁ establishment of large-scale environments favorable for MJO initiation, and S₂ MJO convective outbreaks under the S₁. The S₁ is often explained as “preconditioning” via MJO-scale anomalous horizontal moisture advection (e.g., Kiranmayi & Maloney, 2011; Zhao et al., 2013) and gradual shallow-to-deep pre-moistening called the “discharge-recharge mechanism” (e.g., Benedict & Randell, 2007; Bladé & Hartmann, 1993). Presumably, these processes commonly help MJO convective organization by promoting moisture accumulation.
It is non-trivial when MJO convection is triggered during the preconditioning, however. For instance, Xu and Rutledge (2016) showed that the transition into deep convection is sometimes more rapid than the prediction from the discharge-recharge mechanism. To fill this deficiency in thermodynamic-driven processes, we should scrutinize dynamic variations as external forcing at the $S_z$. Specifically, equatorially circumnavigating Kelvin waves (e.g., Chen & Zhang, 2019; Powell & Houze, 2013; Seo & Kim, 2003) and extratropical disturbances (e.g., Hsu et al., 1990; Ray & Zhang, 2010; Gahtan & Roundy, 2019) can trigger MJO convection by inducing upward motions directly, although it is still debated how robust and plausible the proposed processes are.

This paper focuses on the $S_z$, particularly dynamic roles of mixed Rossby-gravity waves (MRGs) in triggering of MJO convection, which is motivated by several observational studies (Straub & Kiladis, 2003; Takasuka & Satoh, 2020; Takasuka et al., 2019; Yang & Ingersoll, 2011; Yasunaga et al., 2010). Field observations clearly detected MRGs enhanced in the mid-to-upper troposphere during the MJO-suppressed phase (Takasuka et al., 2019; Yasunaga et al., 2010), which supports a notion that those MRGs determine the timing of MJO convective outbreaks in the IO through rapid moistening and/or the development of low-level convergence. Moreover, some previous studies showed that eastward group velocity of MRGs assists the start of MJO propagation (Takasuka & Satoh, 2020; Takasuka et al., 2019; Yang & Ingersoll, 2011).

The aforementioned findings imply that mid-to-upper-tropospheric MRGs may be sometimes influential precursors of MJO initiation. A question here is how upper-tropospheric MRGs finally initiate MJO convection, which is more likely to be affected by lower-tropospheric moisture fields. Takasuka and Satoh (2020) statistically suggested that upper-tropospheric MRG energy input by more diabatic heating associated with MRG–convection coupling results in downward dispersion of MRGs and the formation of low-level MRG wave packets leading to MJO initiation. However, because upper-tropospheric diabatic heating is rooted in lower-tropospheric moisture/wind variations, diabatic processes may not be a primary trigger for low-level MRGs stemming from the upper troposphere. Hence, it is worth examining whether, as an intrinsic mechanism for MJO initiation in which amplification of upper-tropospheric MRGs is involved, there exists a process more in line with upper-tropospheric dynamics.

In this regard, we shed light on the dry interaction between upper-tropospheric MRGs and a sharp downward branch of the Walker circulation (WC) above the western IO (WIO), which is referred to as the “Wall” of WC by Kohyama et al. (2021). Because this “Wall” climatologically forces upper-tropospheric zonal convergence in easterlies over the IO, MRGs approaching there may be amplified by wave accumulation (Hoskins & Yang, 2016) and then be dispersed downward and eastward. Motivated by this insight, we aim to verify the possibility that dry MRG dynamics can play an essential role in MJO initiation, based on simple model simulations and observational data analyses. A possible role of circumnavigating Kelvin waves in the MJO–MRG relationship is also discussed.

2. Data and Model Descriptions

2.1. Observational Data

To provide observational evidence for our hypothesis, we analyze “MJO2” event initiated in mid-November 2011 during a field campaign CINDY2011 (Yoneyama et al., 2013). We use 3-hourly radiosonde observations at Gan Island (0.7°S, 73.2°E), 6-hourly atmospheric fields from ERA-Interim (Dee et al., 2011) with 27 vertical layers spanning 1,000–10 hPa, and 6-hourly rainfall data from the Global Satellite Mapping of Precipitation (GSMaP; Okamoto et al., 2005). A horizontal grid interval of the ERA-Interim (GSMaP) is 0.5° (0.1°). The ERA-Interim data and others covered the entire period of October/November and November 2011, respectively. Note that the boreal-winter (November–March) climatology used in Section 4 is derived from the period of 1979–2012.

Anomalies are calculated by subtracting the mean during the data period. To capture MRG variations, we filter 6-hourly anomalies for westward-propagating wavenumbers and periods of 3.5–8 days (cf. Section 3), using fast Fourier transforms in space and a 101-point Lanczos filter in time (Duchon, 1979).
2.2. Simple Dry Model

Based on Stechmann et al. (2008), a simple dry model with the barotropic and the first and second baroclinic modes for the vertical depth of $H = 16$ km is constructed on the equatorial $\beta$-plane. Note that these three vertical modes can capture the main structure of equatorial waves (e.g., Haertel & Kiladis, 2004; Takayabu et al., 1996). The model equations are

$$\frac{\partial \zeta_0}{\partial t} + v_0 = -\nabla \cdot D_{u_0}(u_0, u_1, u_2) - \frac{\zeta_0}{\tau_u} - K_u \nabla^4 \zeta_0$$

(1)

$$\frac{\partial u_j}{\partial t} + y u_j - \nabla \theta_j = -D_{u_j}(u_0, u_1, u_2) - \frac{u_j}{\tau_u} - K_u \nabla^4 u_j \quad \text{(for } j = 1, 2)$$

(2)

$$\frac{\partial \theta_1}{\partial t} - \nabla \cdot u_1 = -D_{\theta_1}(u_0, u_1, u_2, \theta_1, \theta_2) - \frac{\theta_1}{\tau_\theta} + S_{\theta_1} - K_\theta \nabla^4 \theta_1$$

(3)

$$\frac{\partial \theta_2}{\partial t} - \frac{1}{4} \nabla \cdot u_2 = -D_{\theta_2}(u_0, u_1, \theta_1, \theta_2) - \frac{\theta_2}{\tau_\theta} + S_{\theta_2} - K_\theta \nabla^4 \theta_2$$

(4)

where $\mathbf{u} = (u, v)^T$ is the horizontal wind vector; $\mathbf{u}_j = (-\nu, \nu)^T$; $\zeta$ is relative vorticity; $\tau_u$ ($\tau_\theta$) is the time scale of damping (cooling) for $\mathbf{u}$ ($\theta$); and $S_{\theta}$ is the heat source. Subscripts $j$ for prognostic variables represent the barotropic ($j = 0$) and first and second baroclinic modes ($j = 1, 2$), and the nonlinear advection terms $D_{u_j, \theta_j}$ are given by

$$D_{u_0} = \frac{2}{3} \sum_{j=0}^{2} \mathbf{u}_j \cdot \nabla \mathbf{u}_j + \frac{2}{3} \sum_{j=0}^{2} (\nabla \cdot \mathbf{u}_j) \mathbf{u}_j$$

$$D_{u_1} = \mathbf{u}_0 \cdot \nabla u_1 + \mathbf{u}_1 \cdot \nabla u_0 + \frac{1}{\sqrt{2}} \left[ \mathbf{u}_1 \cdot \nabla u_2 + u_2 \cdot \nabla u_1 + 2(\nabla \cdot u_1) u_2 + \frac{1}{2} (\nabla \cdot u_2) u_1 \right]$$

$$D_{u_2} = \mathbf{u}_0 \cdot \nabla u_2 + u_2 \cdot \nabla u_0 + \frac{1}{\sqrt{2}} [u_1 \cdot \nabla u_2 - (\nabla \cdot u_1) u_2]$$

(5)

$$D_{\theta_1} = \mathbf{u}_0 \cdot \nabla \theta_1 + \frac{1}{\sqrt{2}} \left[ 2u_1 \cdot \nabla \theta_2 - u_2 \cdot \nabla \theta_1 + 4(\nabla \cdot u_1) \theta_2 - \frac{1}{2} (\nabla \cdot u_2) \theta_1 \right]$$

$$D_{\theta_2} = \mathbf{u}_0 \cdot \nabla \theta_2 + \frac{1}{2\sqrt{2}} [u_1 \cdot \nabla \theta_1 - (\nabla \cdot u_1) \theta_1]$$

Equations 1–4 start with the three-dimensional Boussinesq system (Majda, 2003), and they have been nondimensionalized by the scaling used in Stechmann et al. (2008). The derivation of Equations 1–4 is provided in the supporting information (Text S1).

Solutions to the present model are numerically obtained for specific $S_{\theta_1, 2}$ distributions and initial conditions given to examine the interaction between WC and MRGs (see Section 4.1 for details). We assume a zonally periodic meridionally bounded channel of which the zonal and meridional extent is 40,000 km (nearly the circumference along the equator) and 8,000 km, respectively. In all simulations, a grid spacing of 100 km on the Arakawa C-grid and a time step of 15 min for the third-order Runge-Kutta scheme are used. For the fourth-order horizontal diffusion (the damping/cooling) term in Equations 1–4, we adopt $K_u = K_\theta = 1.6 \times 10^{14} \text{ m}^2 \text{ s}^{-1} (\tau_u = \tau_\theta = 20 \text{ days})$.

3. Observational Evidence of MRG Variations Leading to MJO Initiation

“MJO2” event during CINDY2011, initiated in the WIO around November 17 (see Figures 1c–1e for 10°N–10°S rainfall variations in the time–longitude sections), stems from amplification of upper-tropospheric MRGs. In Figure 1a, the wavelet analysis (Torrence & Compo, 1998) for radiosonde-derived meridional winds at Gan highlights significant 4–5.5-day period variations at 300–200 hPa during November 5–12 (shading), after enhanced lower-tropospheric variations in the 6–8-days cycle (contours). These wind variations, detected from a 3.5–8-days-filtered data, are associated with cross-equatorial circulations with equatorially symmetric meridional wind signals (Figure 1b), indicating the robust MRG structure. This amplification of upper-tropospheric MRGs is followed by re-intensification of lower-tropospheric MRGs in the end of November during the MJO-active phase (Figures 1a and 1b).
The aforementioned fact is reinforced by the time–longitude diagrams of equatorial MRG-filtered meridional wind anomalies in the upper/lower troposphere and non-filtered precipitation field (Figures 1c and 1d). The eastward propagation of MJO2 precipitation in the IO appears to collocate with the eastward formation of lower-tropospheric MRG wave packets beginning with northerlies in 45°–60°E (Figure 1c). In fact, low-level MRG circulations successively trigger MJO convection from the WIO (Text S2 and Figure S1), consistent with the view that MRGs can actively contribute to MJO convective initiation (Takasuka & Satoh, 2020; Takasuka et al., 2019). Before this situation, around November 10, upper-tropospheric MRG variations begin to strengthen over the WIO in conjunction with the slowdown of their westward propagation (Figure 1d; magenta lines), which slightly precedes the development of the lower-tropospheric MRG wave packets in 45°–60°E. This evolution is also reconfirmed from the MRG-related eddy kinetic energy (EKE) field, defined by $K' = \frac{(u')^2 + (v')^2}{2}$ where primes denote MRG-filtered values; the positive tendency and subsequent accumulation of upper-level EKE is evidently observed over the WIO before MJO2 initiation (Figure 1e).
4. Mechanism

Based on the analyses in Section 3, we raise two questions: Why are upper-tropospheric MRGs amplified in the WIO? How are low-level MRG wave packets leading to MJO initiation formed? As for the former question, the amplifying upper-tropospheric MRGs with the slowdown of their phase propagation are reminiscent of the interaction with zonally varying background flows (Hoskins & Yang, 2016). Inspired by this idea, we deductively examine the above questions with simple dry model experiments by focusing on a role of WC, which has the Wall (i.e., the zonally narrow downward branch of WC) above the WIO (Kohyama et al., 2021). In parallel, we show that the presented mechanism is applicable to the MJO2 event.

4.1. Relationship Between the Walker Circulation and MRGs

First, WC in the model is obtained as the steady-state response to the time-invariant heat source \( \dot{S}_\alpha \) at \( z = 8 \) km (pink) and observed apparent heating \( \overline{Q}_\alpha \) at 400 hPa (blue). (b)–(d) Zonal-height sections of vertical velocity (shading), zonal-vertical winds (vectors), and zonal convergence (contours) for (b) the spin-up model simulation at day 200, (c) boreal-winter mean, and (d) 11 days running mean fields during November 5–15. All fields are subtracted from their zonal mean, and averaged over \( \gamma = \pm 1.050 \) km (10°S–10°N) range for the model (observation). Contour interval in (b and d) [(c)] is 1.0 [0.5] \( \times 10^{-5} \) s, with zero contours bolded. Contours below 750 hPa in (c and d) are masked for visibility, and vertical velocity for vectors in (b), and (c and d) is multiplied by 1,000 and 400, respectively.

Figure 2. (a) Zonal distributions of the heat source for the first baroclinic mode in the model \( \dot{S}_\alpha \) at \( z = 8 \) km (pink) and observed apparent heating \( \overline{Q}_\alpha \) at 400 hPa (blue). (b)–(d) Zonal-height sections of vertical velocity (shading), zonal-vertical winds (vectors), and zonal convergence (contours) for (b) the spin-up model simulation at day 200, (c) boreal-winter mean, and (d) 11 days running mean fields during November 5–15. All fields are subtracted from their zonal mean, and averaged over \( \gamma = \pm 1.050 \) km (10°S–10°N) range for the model (observation). Contour interval in (b and d) [(c)] is 1.0 [0.5] \( \times 10^{-5} \) s, with zero contours bolded. Contours below 750 hPa in (c and d) are masked for visibility, and vertical velocity for vectors in (b), and (c and d) is multiplied by 1,000 and 400, respectively.
These \( S_{\theta_0} \) produce the realistic WC after the 200 days from the state of rest, as recognized by a comparison of WC for the model and observed boreal-winter mean (Figures 2b and 2c); the Wall and associated upper-tropospheric zonal convergence over the WIO are reproduced. As expected, the same features as climatology are also realized in the 11 days running mean zonal-vertical circulations before MJO2 initiation (during November 5–15; Figure 2d), except for stronger zonal convergence than for the climatology (or the model), which will be discussed later.

Under the simulated WC, we examine how upper-tropospheric MRGs as observed before MJO initiation evolve. Referring to observations (Figure 1d and Figure S4), we set the initial MRG structure as the zonal wavenumber-8 mode confined in \( 7500 \leq x \leq 9000 \) km (i.e., the eastern side of the Wall) with maximum amplitudes at the model top, in the manner of Aiyyer and Molinari (2003) (see Text S3 and Figure S2 for details). This setting is somewhat arbitrary, because what is the ultimate source of such MRGs is still unclear. As one speculation, meridionally asymmetric external environments of the IO (e.g., SST) might be likely to induce equatorially anti-symmetric modes when the basin-scale convective variability is realized. This idea will be tested in our future work. Here, from the initial condition prepared by superimposing the derived MRG field onto the steady state obtained from a 200 days spin-up integration, we run the model for 30 days.

Figure 3. (a and b) Time-longitude diagrams of 24 h running mean (MRG-filtered) meridional wind anomalies in the upper/lower troposphere (shading/contours) for the model (MJO2). Contours denote \( \pm 0.2, \pm 0.3, \ldots \ (\pm 1.0, \pm 2.0, \ldots) \) m/s with negative values dashed. Purple broken-line squares enclose the domains used to compute EKE budget shown in Figure 4, and red and blue broken-line arrows indicate group propagation of upper- and lower-tropospheric MRG wave packets, respectively. Magenta lines in (b) present the phase speed change of MRGs. (c and d) Zonal distributions of upper-/lower-tropospheric zonal winds (blue/pink) and upper-tropospheric zonal convergence (green) for the simulation-period mean in the model (11 days running mean for MJO2). Upper-tropospheric (Lower-tropospheric) fields for MJO2 are computed during November 5–15 (17–27). Broken lines in (d) denote the boreal-winter mean. (e) Time-longitude diagram of 5 days running mean upper-tropospheric zonal wind anomalies for MJO2. Pink broken-line square enclose the domain into which the westerly phase of a large-scale Kelvin wave intrude. All fields in (a), (b) and (c)-(e) are averaged over \( y = \pm 550 \) km (7.5°S–7.5°N) range for the model (MJO2), respectively. The upper and lower troposphere for the model (MJO2) are defined as the 13.2–16 km and 0–2.8 km (300–200 hPa and 1000–800 hPa) layer, respectively.
easterlies (Figure 3c). Along with this contraction, upper-tropospheric MRGs are gradually amplified in the Wall (4500 ≤ x ≤ 5500 km) until around day 15 when they begin to exhibit eastward group velocity (red arrow), and then lower-tropospheric MRG wave packets are radiated eastward (blue arrow).

In Figures 3b and 3d, which are the same as Figures 3a and 3c but for the observed MJO2, the processes predicted by the model are similarly detected. After November 5, upper-tropospheric MRGs propagating westward with small positive group velocity are decelerated (magenta lines) and amplified in 45°–60°E, where the zonal convergence associated with the Wall is realized. Then, lower-tropospheric MRG wave packets moving eastward are established, which characterizes MJO2 initiation.

Despite much consistency between the model and MJO2, there are some noteworthy differences. One is faster group velocity of the lower-tropospheric MRGs in the model (Figures 3a and 3b). This is attributed to the doppler shift by stronger background low-level westerlies (Figures 3c and 3d) and deeper equivalent depth in the dry model. The latter reflects the limitation that dry dynamics cannot represent wave–convective coupling effects that are important after MJO initiation.

Another difference is the stronger upper-tropospheric background zonal convergence around the Wall before MJO2 initiation (Figures 3c and 3d). This is because the observed background WC for MRGs are contributed by not only the climatology but also large-scale circumnavigating Kelvin waves with their evolution slower than MRGs. In fact, upper-tropospheric westerlies associated with circumnavigating Kelvin waves intrude into the WIO (Figure 3c; pink square), in agreement with the stronger background westerlies to the west of 50°E than the boreal-winter mean (Figure 3d). This process, which is not incorporated in the model, promotes convergence with climatological upper-level easterlies. Considering that zonal convergence can amplify MRGs (see Section 4.2), upper-tropospheric circumnavigating Kelvin waves could serve as a catalyst of MRG-induced MJO initiation.

4.2. Amplification of Upper-Tropospheric MRGs and Its Impacts on the Lower Troposphere

To reveal why upper-tropospheric MRGs are amplified around the Wall and then lower-tropospheric MRG wave packets are formed there, we conduct the EKE budget analysis. The budget equation for the model is

\[
\frac{\partial \mathcal{K}}{\partial t} = -\mathbf{u} \cdot (\nabla \cdot \mathbf{u}) - \mathbf{v} \cdot \nabla \mathcal{K} - \nabla \cdot (\mathbf{v} \nabla \mathcal{K}) + \nabla \cdot (\mathbf{v} \mathcal{W}) + \left( \mathcal{W} \mathcal{T} \right) + \left( \mathcal{R} \right) \tag{6}
\]

where \( \mathbf{v} \) is the three-dimensional wind vector; \( \mathbf{w} \) is vertical velocity; and overbars (primes) denote 11 days running mean (deviations from the mean of the 30 days simulation). For the ERA-Interim, primes denote MRG-filtered values, and \( P_t K_x \) and \( G K_x \) terms are replaced with \( -(R_0 \rho \omega T^2) \) and \( \nabla \cdot (\mathbf{v} \mathcal{T}) \), respectively, where \( \omega \) is vertical p-velocity; \( T \) is temperature; \( \Phi \) is geopotential; and \( R \) is the gas constant. Note that real sources/sinks of EKE are \( K_u K_v \) and \( P_t K_v \).

Figures 4a and 4d compare upper-tropospheric EKE budget terms averaged in the time-area domain where MRG amplification occurs for the model and MJO2 (see broken-line squares in Figures 3a and 3b). This comparison shows physical consistency with each other; upper-tropospheric MRGs are amplified by EKE advection by background flows \( (A_u K_v) \) and the barotropic conversion from the background \( (K_u K_v) \). In Figures 4b and 4e, the decomposition of these terms,

\[
(A_u K_v) = -\bar{u} \frac{\partial \mathcal{K}'}{\partial x} - \bar{v} \frac{\partial \mathcal{K}'}{\partial y} - \bar{w} \frac{\partial \mathcal{K}'}{\partial z} \tag{7}
\]

\[
(K_u K_v) = -\bar{u} \bar{v} \frac{\partial \mathcal{K}'}{\partial x} + \bar{u} \bar{w} \frac{\partial \mathcal{K}'}{\partial y} - \bar{u} \bar{w} \frac{\partial \mathcal{K}'}{\partial z} \tag{8}
\]

reveals that \( A_u K_v \) and \( K_u K_v \) processes are dominantly contributed by \(-\bar{u} \bar{v} \mathcal{T} \) and \(-\bar{u}^2 \mathcal{T} \), respectively. This result ensures the following interpretation for upper-tropospheric MRG amplification: upper-level easterlies of WC into the Wall efficiently advects MRG energy from the east of the Wall, and advected energy is further amplified by wave accumulation due to zonal convergence arising from the Wall. Because westward-propagating MRGs (with typical group velocity ~5 m/s) are accumulated for their
positive ground group velocity in zonal convergence (Hoskins & Yang, 2016), the region near the Wall with \( \vec{u} > -5 \) m/s is indeed appropriate for MRG accumulation (cf. Figure 3).

Also in the lower-troposphere, the EKE tendency is positive for both the model and MJO2 (Figures 4c and 4f), corresponding to the formation of low-level MRG wave packets. For the model (Figure 4c), this positive tendency almost originates from the EKE redistribution via potential eddy flux convergence (\( G_K \)). Because EKE source here is only upper-tropospheric \( K_u K_v \) (Figures 4a and 4c), the lower-tropospheric EKE is brought by the energy dispersion from the upper troposphere. Positive lower-tropospheric \( G_K \) with positive upper-tropospheric \( K_u K_v \) is also observed for MJO2 (Figures 4d and 4f), supporting a notion that process found in the model operates before MJO2 initiation, despite the difference in lower-tropospheric \( K_u K_v \) contributions.

The downward impacts of amplification of upper-tropospheric MRGs are qualitatively inferred from the vertically eastward-tilted MRG structure (Figure S3) and equatorial zonal-height sections of \(-\frac{\partial}{\partial y} \left( \vec{u} \vec{v} \right)\) and \( G_K \) (Figures 4g and 4h). For the model (Figure 4g), MRG-related EKE is accumulated especially in the inner Wall (4500 ≤ x ≤ 5500 km) in the upper troposphere, and as indicated by positive \( G_K \) below it, the accumulated EKE is redistributed to the mid-to-lower troposphere. The physical essence of this situation holds true for MJO2 (Figure 4h), despite not being a very strong resemblance between Figures 4g and 4h.
EKE redistribution by $GK_e$ occurs below the positive $-u^2(\frac{\partial u}{\partial x})$ area within the Wall ($40^{\circ}$–$60^{\circ}$E), especially to the west of $45^{\circ}$E and around $60^{\circ}$E.

To make the above view more compelling for observation, we conduct 10-days ray tracing of MRGs from around $49^{\circ}$E, 300 hPa ($z \sim 9680$ m), and November 10, where and when MRG amplification is clearly observed (Figures S3b and S4). The initial zonal and vertical wavelength ($\lambda_z$ and $\lambda_z$) for ray tracing is roughly estimated as $\lambda_z \sim 47^{\circ}$ and $\lambda_z \sim 20$ km from the vertical structure (Figure S3b). For those parameters and $\bar{u} = -5.5$ m/s, the MRG dispersion relation predicts ground zonal phase speed $c_{pz} \sim -17$ m/s, consistent with MRGs propagating into the WIO (Figure S4). Practically, $\lambda_z$ is difficult to be identified from the vertically coarse data, so initial $\lambda_z$ is determined by the MRG dispersion relation with initial $\lambda_z$ and $c_{pz} = -17$ m/s.

This ray tracing reconfirms the downward-eastward energy dispersion of amplified upper-tropospheric MRGs. In Figure 4h, the rays for 45 initial conditions considering their estimation uncertainties (see Text S4 for method details) indicate that a fraction of rays reach the mid-to-lower troposphere in $50^{\circ}$–$70^{\circ}$E after “reflection” in the Wall, although others go through the Wall westward (as indicated by $GK_e$ distributions).

### 5. Summary and Discussion

In this study, we have presented a new pathway to MJO initiation that stems from dry upper-tropospheric westward-propagating MRGs above the IO. This is inspired by initiation processes of the “MJO2” event during CINDY2011, in which upper-tropospheric MRG amplification in the WIO is followed by MJO2 initiation (Figure 1). Here we hypothesize that the interaction between MRGs and the Walker circulation (WC) is potentially important.

To test our hypothesis, we perform numerical simulations using a simple dry model with three vertical modes, comparing the model output with observations for MJO2. The model captures the essence of the boreal-winter mean WC above the IO: upper-level zonal convergence in mean easterlies blowing into the WIO, where the “Wall” (i.e., the zonally narrow downward branch of WC) exists (Figures 2b and 2c). In the model with this idealized WC, upper-tropospheric MRGs propagating into the Wall are amplified in the inner region of the Wall. Then, lower-tropospheric MRG wave packets start to propagate eastward (Figures 3a and 3c), resembling the processes of MJO2 initiation triggered by low-level MRG wave packets with eastward group velocity (Figures 3b and 3d).

The energetics for this MRG evolution is discussed by both the model experiment and observations (Figure 4). The initial amplification of upper-tropospheric MRGs in the Wall results from MRG energy advection to the Wall and wave accumulation due to upper-level easterlies of WC and their zonal convergence arising from the Wall. Subsequently, the eastward-downward dispersion of the amplified upper-level MRG energy is activated, which forms lower-tropospheric MRG wave packets leading to MJO initiation.

A difference of WC between the model and MJO2 (Figures 2b–2d) has implication that upper-tropospheric circumnavigating Kelvin waves make the presented mechanism more efficient by modulating background WC additionally. For MJO2, upper-level zonal convergence in the Wall are enhanced by cooperation between the westerly phase of Kelvin waves propagating into the WIO and climatological easterlies of WC above the IO (Figure 3e), which promotes MRG-wave accumulation. In addition, upper-tropospheric Kelvin-wave westerly anomalies help the realization of positive ground group velocity of MRGs by weakening upper-tropospheric mean easterlies, which is advantageous to triggering the wave accumulation (Hoskins & Yang, 2016). Furthermore, the easterly phase of Kelvin waves before the westerly phase can enhance westward advection of MRG energy into the WIO. For these reasons, equatorial circumnavigation of Kelvin waves assists MRG-induced MJO initiation cooperatively with the climatological WC.

The idea proposed in this study for MJO initiation does not require moist processes at all, which provides several debatable topics. First, we may reconsider roles of diabatic processes in the similar MRG-related mechanism suggested by Takasuka and Satoh (2020) and Takasuka et al. (2019). A possible interpretation for this is that dry dynamics are sufficient for an initial trigger of amplification of upper-tropospheric MRGs, although diabatic heating can accelerate and/or maintain MRG amplification in a later stage when MRG-convection coupling becomes evident. Second, our idea does not necessarily contradict with the preexisting hypotheses that put emphasis on moisture variations (e.g., Benedict & Randall, 2007; Zhao et al., 2013),
because we have addressed MJO initiation in terms of convective triggering by gravity wave dynamics (e.g., Tulich & Mapes, 2008), assuming a favorable environment for organized convection regulated by moisture fields. Nevertheless, if dry MRG dynamics by itself can determine the timing of MJO initiation, it would be misleading to emphasize only the moisture variations for understanding MJO initiation. Because a simple dynamical model theoretically predicts the dry interaction between upper-tropospheric MRGs and WC as observed for a single MJO event, the next step is to examine its robustness and relationship with moist processes statistically for multiple cases.

Data Availability Statement
The ERA-Interim data are available online (https://www.ecmwf.int/en/forecasts/datasets/archive-datasets/reanalysis-datasets/era-interim); GSMaP product is distributed online (https://sharaku.eorc.jaxa.jp/GSMaP/index.html); and radiosonde-derived data during CINDY2011 are archived at https://data.eol.ucar.edu/dataset/347.008.

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