Dynamical Roles of Mixed Rossby–Gravity Waves in Driving Convective Initiation and Propagation of the Madden–Julian Oscillation: General Views

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ABSTRACT: Motivated by the previous case study, this work shows that dynamical variations of mixed Rossby–gravity waves with tropical depression–type circulations (MRGTDs) are possible drivers of convective initiation and propagation of the Madden–Julian oscillation (MJO) by performing statistical analysis. MJO events initiated in the Indian Ocean (IO) in boreal winter are objectively identified solely using outgoing longwave radiation data. The lagged-composite analysis of detected MJO events demonstrates that MJO convection is initiated in the southwestern IO (SWIO), where strong MRGTD–convection coupling is statistically found. Further classification of MJO cases in terms of intraseasonal convection and MRGTD activities in the SWIO suggests that 26 of 47 cases are related to more enhanced MRGTDs, although they can also be secondarily affected by Kelvin waves. For those MRGTD-enhanced MJO events, MJO convective initiation is primarily triggered by low-level MRGTD circulations that develop via the enhancement of downward energy dispersion in accordance with upper-tropospheric baroclinic conversion. This is supported by the modulation of MRGTD structure associated with zonal wave contraction due to upper-tropospheric zonal convergence, and plentiful moisture advected into the western IO. Following this MRGTD-induced MJO triggering and midtropospheric premoistening in the IO contributed by MRGTD shallow circulations as well as intraseasonal winds during the MJO-suppressed phase, low-level MRGTD winds with eastward group velocity successively trigger convection to the east, which helps MJO convective propagation over the IO. The interannual atmospheric variability may affect whether the presented MRGTD-related processes are effective or not.

KEYWORDS: Madden–Julian oscillation; Waves, atmospheric; Intraseasonal variability

1. Introduction

The Madden–Julian oscillation (MJO) (Madden and Julian 1971, 1972) is known to be the most prominent convective disturbances in the tropics. It is characterized by slow eastward propagation of convective envelopes with $O(10^3)$ -km spatial scale over the Indo-Pacific warm pool sector on an intraseasonal time scale (e.g., Hendon and Salby 1994; Kiladis et al. 2005), and it appears not to be analogous to classical equatorial waves (Matsuno 1966; Takayabu 1994; Wheeler and Kiladis 1999). Because the MJO has profound impacts on weather and climate phenomena (Zhang 2013), gaining a deeper understanding of the mechanics of MJO initiation and propagation is an important goal in tropical meteorology. Although extensive work over the past several decades has improved our knowledge of the MJO (Zhang 2005; Lau and Waliser 2012), its intrinsic mechanism remains elusive. In this context, many general circulation models (GCMs) still struggle with simulating MJO convection with appropriate amplitude, propagation characteristics, and frequency of occurrence (e.g., Hung et al. 2013; Jiang et al. 2015; Ahn et al. 2017; Ling et al. 2019).

While the *moisture* evolution under the weak temperature gradient approximation (Sobel et al. 2001) and the quasi-equilibrium framework (e.g., Betts and Miller 1986; Emanuel et al. 1994) at the MJO scale could explain the robust properties of MJO convection (e.g., Kiranmayi and Maloney 2011; Sobel and Maloney 2013; Adames and Kim 2016), it is natural to consider about the contributions of high-frequency wave *dynamics* to the MJO if we keep in mind the MJO hierarchical structure (e.g., Nakazawa 1988; Hendon and Liebmann 1994; Kikuchi and Wang 2010). One may expect that, as suggested by Emanuel et al. (1994), high-frequency wave modes can be damped at the quasi-equilibrium state for the MJO scale because a finite time that convective adjustment to the quasi-equilibrium takes leads to negative correlations of heating and temperature fields associated with first-mode baroclinic waves. In this idea, however, the assumption of the vertical wave structure is too simple compared to an observational fact that some waves have the significant second baroclinic mode (e.g., Kiladis et al. 2009). Related to this, Kuang (2008) discusses the destabilization of equatorial waves using a simple model that incorporates the two-vertical mode under the quasi-equilibrium concept.

In fact, high-frequency waves modulate part of MJO convection via wave-induced convective triggering (Masunaga et al. 2006; Yang and Ingersoll 2013, 2014) and wave energy dispersion (Straub and Kiladis 2003; Yang and Ingersoll 2011).

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Masunaga et al. (2006) suggest that consecutive Kelvin waves interfered with by Rossby waves help to trigger and maintain the MJO envelope. Yang and Ingersoll (2013, 2014) have proposed a theory that the interference of westward and eastward inertia-gravity waves can realize slow MJO propagation using a 2D shallow-water system that implements the effects of triggered convection. Straub and Kiladis (2003) argue that MJO convection during boreal summer can be a packet of convectively coupled mixed Rossby-gravity (MRG) waves-tropical depression (TD)-type disturbances (MRGTDs). This argument is a valid hypothesis according to idealized numerical experiments with a dry GCM (Yang and Ingersoll 2011) and is consistent with the fact that the MRG group velocity is almost equal to the MJO propagation speed (Wheeler and Kiladis 1999).

Meanwhile, there is a question regarding the interaction between the MJO and high-frequency equatorial waves: Do particular wave modes have upscale effects on the MJO? Using a multiyear outgoing longwave radiation (OLR) dataset, Yang and Ingersoll (2011) report statistically significant but small contributions of MRGs to the MJO envelope. More generally, Dias et al. (2017) indicate that overall, rather than specific, wave activities are enhanced during an active-MJO phase in terms of convection. However, if we focus only on convective signals, ambiguities are introduced to the understanding of the cross-scale interaction because the MJO convective complex can be aggregation of convection that is triggered externally by high-frequency dynamics (e.g., synoptic-scale waves) and spontaneously develop in response to thermodynamical (e.g., moisture) variations due to convective adjustment to the quasi-equilibrium. Alternatively, it may be better to clarify the roles of a specific wave type in the MJO with an emphasis on wave-induced dynamical aspects such as convective triggering and moisture transport.

From this standpoint, our interest in the present study is the MJO-MRGTD interaction, which is motivated by the previous study by the authors (Takasuka et al. 2019, hereafter TSY19). TSY19 suggest that MJO convection during the Years of the Maritime Continent (YMC)-Sumatra 2017 campaign was triggered in the southwestern Indian Ocean (SWIO) by MRG-convection coupling associated with a similar analogy to MRG-TD transition, and propagated over the Indian Ocean (IO) in accordance with successive convective triggering by MRG-related meridional winds with eastward energy dispersion. TSY19 further confirmed midtropospheric moistening associated with MRG shallow circulations prior to an MJO convective outbreak.

A similar perspective—that MRGTD dynamics affect moisture modulation and convective triggering that are responsible for MJO realization in the IO—has been proposed for cases in different field campaigns (Yasunaga et al. 2010; Kerns and Chen 2014; Chen et al. 2015; Muraleedharan et al. 2015) and an idealized numerical study (Takasuka et al. 2018). Yasunaga et al. (2010), who analyzed observational data from the *Mirai* Indian Ocean cruise for the Study of the MJO-convection Onset (Yoneyama et al. 2008), found midtropospheric meridional wind variations associated with MRGs and related moisture resurgence before the MJO initiation. During

the Dynamics of the MJO (DYNAMO) campaign (Yoneyama et al. 2013), MRGs and/or TD-type disturbances helped the organization of MJO convection through dry air intrusion (Kerns and Chen 2014) and/or vertical and horizontal moisture advection (Chen et al. 2015; Muraleedharan et al. 2015). Furthermore, the importance of midtropospheric moistening due to MRG shallow circulations before MJO initiation was shown in long-term aquaplanet experiments using the Nonhydrostatic Icosahedral Atmospheric Model (Takasuka et al. 2018).

Because the MJO-MRGTD interaction presented by TSY19 and others is derived from limited case studies or idealized experiments, it is unknown whether MRGTDs are actually a driving force for the MJO in general. In this paper, using the datasets and analysis methods described in section 2, we aim to present that convective triggering and moistening associated with MRGTDs is one possible way to MJO initiation and propagation in more general context, even though it is not necessarily applicable to all MJO events. We begin by detecting MJO events initiated in the IO during boreal winters of 1982–2012 and specifying the MRGTD structure in the IO statistically (section 3). We then show that the MJO cases triggered by MRGTDs in the SWIO are not rare; for those cases, we clarify the robust contributions of MRGTD dynamics to dynamical and moistening processes that lead to MJO convective initiation and propagation over the IO by applying lagged-composite and regression analyses (section 4). In section 5, we discuss factors in large-scale environments that could potentially make a difference to MRGTD-enhanced and other types of MJO initiation. Summary and concluding discussion comprise section 6.

2. Data and methodology

a. Data

We used the interpolated daily OLR from the National Oceanic and Atmospheric Administration (NOAA) polar-orbiting satellite (Liebmann and Smith 1996) and 6-hourly atmospheric fields from ERA-Interim (Dee et al. 2011). The OLR dataset used to identify MJO events and deep convective activities in the tropics has a horizontal resolution of 2.5° latitude $\times$ 2.5° longitude for the 35-yr time period from 1979 to 2013. The ERA-Interim data were used to investigate dynamical and thermodynamical variations from January 1979 to December 2012. Their horizontal spatial resolution is 1.5° latitude $\times$ 1.5° longitude and they cover 27 vertical layers from 1000 to 100 hPa. We used the daily averaged 3D field variables of horizontal wind components [$\mathbf{V}_h = (u, v)$], vertical p velocity (ω), specific humidity (q), temperature (T), and geopotential (Φ), as well as surface pressure (p_s) to derive column-integrated (surface–100 hPa) water vapor (CWV). ERA-Interim OLR at 0000 and 1200 UTC was also used to filter out high-frequency convective variations with the period around 2–3 days, which cannot be completely captured by the daily data alone. In this study, we analyzed the atmospheric fields in boreal winter [November–April (NDJFMA)].

b. Analysis methods

We first detected MJO events initiated over the IO in December–March (DJFM) of 1982–2012, which corresponds to the post–geostationary satellite era and season with robust MJO activity, in terms of convective fields by tracking amplitudes and phases in an MJO phase space based on the EOF analysis of the intraseasonal OLR anomalies in the tropics (30°S–30°N, all longitudes) in DJFM of 1980–2012. While EOF-based multivariate indices constructed from OLR and 850- and 200-hPa zonal wind anomalies in the equatorial region are widely used to monitor MJO activity (Wheeler and Hendon 2004), they tend to miss MJO-related local convection, especially that seen in MJO initiation, because of their emphasis on the planetary-scale dynamical circulation (Straub 2013). We thus utilized only intraseasonal OLR anomalies to track MJO convection from the MJO initiation stage (cf. Matthews 2008; Kikuchi et al. 2012; Kiladis et al. 2014). The intraseasonal anomalies were defined as follows: the daily OLR anomalies from the climatology were first obtained by subtracting the mean and the first three harmonics of the annual cycle in 1979–2013 and then filtered for the periods of 20–100 days with a 201-point Lanczos filter (Duchon 1979). The detailed procedure of MJO detection is provided in section 3a.

To obtain MRGTD characteristics in the IO, we adopted temporal and cross-spectral analyses and a linear regression technique following Kiladis et al. (2009). In the spectral analysis, detrended time series of westward-only-filtered anomalies of any variables were divided into 180-day segments from 1 November in each year from 1981. For each segment, the time mean was removed and the ends of the series were tapered to zero. Then we applied fast Fourier transforms (FFT) in time to all 31 segments, averaged them, and applied a 1–2–1 filter to the averaged spectra. With the filtering band of MRGTDs in frequency (4–10 days) and wavenumber (–1 or less) domains constructed from the spectral analysis, we extracted composite structure of MRGTDs by regressing any nonfiltered variables at all grid points onto MRGTD-filtered OLR anomalies at a given grid point. The data for this regression analysis covers NDJFMA in 1982–2012. Similar to Kiladis et al. (2009), we assessed the statistical significance of the local regressed fields by applying a two-tailed test for the Fisher transformed correlation coefficients at the 95% level, based on degrees of freedom estimated from the decorrelation time scale in Livezey and Chen (1983). The regression results were scaled to –2.0 standard deviations of MRGTD-filtered OLR for lag 0.

The statistical results for the MJO–MRGTD relation were derived primarily from lagged-composite analysis with reference to the MJO initiation date assigned in MJO detection. Unless otherwise noted, the anomalies in this analysis were defined as 5-day running-mean deviations from 91-day running means of any variables to remove their daily climatology and interannual variability. The statistical significance of the composite fields at any grid points were assessed by the two-tailed Student's *t* test at the 95% or 90% level. We also conducted the lagged-regression analysis to clarify the time evolution of

dynamical fields associated with MRGTDs in selected MJO events (see section 4c).

3. Identification of MJO events and MRGTD structure in the IO

a. Detection of the MJO and its composite feature

As described in section 2b, the EOF analysis for tropical intraseasonal OLR anomalies was employed to construct the MJO phase space used for MJO detection. EOF1 and EOF2 represent large-scale enhanced convection over the Maritime Continent (MC) and the IO, respectively (Fig. 1a), and the corresponding first principal component (PC1) lags the second one (PC2) by about 10 days (not shown). Thus, the phase space constructed from PC1 and PC2 well captures the evolution of MJO convection. In fact, horizontal maps of OLR and 850-hPa horizontal wind anomalies averaged on the days with amplitude $A \geq 0.8$ ($A \equiv \sqrt{PC1^2 + PC2^2}$) in each of the eight phases (Figs. 1b,c) suggest that active MJO convection starts from the SWIO to the central IO at phases 1–3 and propagates eastward to the western Pacific as phases progress, which confirms the availability of the OLR-based MJO index from PC1 and PC2 for tracking MJO activities.

Following the methodology used in Takasuka et al. (2018), we took two steps to detect MJO events. First, a large-scale convective event (LCE) over the IO (10°S–10°N, 60°–90°E) was directly identified using the time series of domain-mean OLR anomalies from the 91-day running mean. We searched the date t_1 , which is the first of five consecutive days for which 5-day running-mean OLR anomalies averaged over the IO (OLR'_{area}) are less than -0.6 standard deviation of OLR'_{area} (σ). We also defined the date t_2 as the first of five consecutive days in which $OLR'_{area} > -0.6\sigma$ after t_1 . If there are any days during $t_1 - t_2$ on which $OLR'_{area} \leq -0.8\sigma$ was satisfied, the first and last of them are those of an LCE period (denoted as D_{start} and D_{end} , respectively). Critical values of σ were subjectively chosen, as they can capture as many plausible LCEs recognized in time–longitude diagrams of equatorial OLR anomalies as possible.

Second, we examined whether the identified LCE was related to the MJO convective initiation by tracking the OLR-based index according to five criteria modified from Suematsu and Miura (2018) and Takasuka et al. (2018). We here focused on the LCEs whose periods include DJFM in 1982–2012. The first criterion (C_1) is $A \geq 0.8$ during an LCE period (i.e., $D_{start} - D_{end}$) in any of phases 1–3 that correspond to MJO initiation over the IO (Fig. 1b). The remaining four criteria are imposed on tracking after the first date when C_1 is satisfied (denoted as D_{init}):

C_2 : $A \geq 0.4$ during tracking.

C_3 : $A \geq 0.8$ at phases 1–3 unless there is a period of $(0.4 \leq A < 0.8)$ within 5 days.

C_4 : No phase is skipped and there is no more than one phase recession.

C_5 : Complete tracking up to phase 5 with C_2 – C_4 satisfied.

Note that if phase recession from phase 1 to phase 8 occurred during tracking and C_1 was newly fulfilled in the LCE period,

FIG. 1. (a) The first two EOF spatial patterns obtained from 20- to 100-day-filtered daily OLR anomalies in 30°S–30°N and at all longitudes. The variance contribution rate of each mode is shown in the top-right corner of each panel. (b) Maps of composite OLR (shading; W m^{-2}) and 850-hPa horizontal wind anomalies (vectors; m s^{-1}) for each MJO phase. Green squares in phases 1–3 denote the Indian Ocean (10°S–10°N, 60°–90°E). The number of days used for the composite in each phase is displayed in the bottom-right corner of each panel. (c) The evolution of PC1 and PC2 during a 31-day period for two LCEs identified as an MJO (red) and a non-MJO event (blue). Each day is plotted with a filled marker. The number plotted every 5 days indicates the day from D_{start} (large filled marker), the star and crossing denote the start and end date of tracking, respectively, and the open circle denotes D_{end} .

D_{init} was updated and another tracking was performed under criteria C_2 – C_5 . The choice of the threshold of A is based on several previous works; Suematsu and Miura (2018) confirmed sufficiently strong convection under $A \geq 0.8$ in the MJO index by Wheeler and Hendon (2004) and Matthews (2008) found that $A \geq 0.4$ in the OLR-based MJO index is enough to track eastward-propagating MJO signals. These references are helpful for maximizing the performance of MJO detection, although our major results are not so sensitive to the thresholds.

In the physical context, C_1 and C_3 guarantee the establishment and maintenance of robust MJO convection in the IO; C_2 and C_4 require the smooth eastward propagation of the MJO structure from the IO to the MC with certain strength; C_5 requires that the MJO completes its propagation over broad spatial scales. We defined LCEs that met criteria C_1 – C_5 as MJO events initiated on D_{init} . If two successive D_{init} had an interval of 20 days or less, the latter D_{init} was regarded as the MJO initiation date to prevent a single MJO event from being counted twice.

Figure 1c shows an example of the tracking for two LCEs classified into an MJO and non-MJO. This MJO satisfies C_1 on

D_{start} (defined as day 0; large star with a filled circle) and succeeds in phase progression until C_5 is met on day 18, so D_{init} of this event was defined as D_{start} . Meanwhile, the non-MJO event misses C_3 on day 14, even though C_1 is satisfied on day 2 (star). In this manner, we detected 47 MJO events used for our analysis.

To examine the initiation processes of 47 MJO cases, we looked at the composite time evolution of OLR, 850-hPa wind, and CWV anomalies from days -9 to 0 (Fig. 2). The composite center (day 0) is D_{init} of each MJO. On day -9 , the anomalously moistened area with easterlies is in the WIO, to the west of suppressed MJO convection, and it is more obvious to the northern side of the equator (Fig. 2a). We recognize enhanced convection over the SWIO on day -3 and the start of its eastward propagation by day 0 (Figs. 2c,d). Notably, the equatorial asymmetry in moisture and convection is detectable before MJO initiation even in composite fields, as in a case study of Chen et al. (2015) and TSY19, which suggests that MRGTDs might potentially contribute generally to MJO realization. More specifically, we expect that active convection

FIG. 2. Maps of composite OLR (shading; W m^{-2}), horizontal wind at 850 hPa (vectors; m s^{-1}), and CWV anomalies (contours; kg m^{-2}) every 3 days from day -9. Contour interval is 1.5 kg m^{-2} , with negative (zero) contours dashed (omitted). Stippling over shaded areas and vectors denote statistical significance at the 95% level.

on day -3 is the result of convective triggering related to strong MRGTD-convection coupling as suggested by TSY19. However, to the best of our knowledge, only Fukutomi and Yasunari (2013) have shown evidence of MRG-like activities with TD-type circulations around this region. Thus, we next sought to clarify the general characteristics of MRGTDs in the IO.

b. MRGTD characteristics in the IO

To extract the representative period of MRGTDs over the IO, if any, in terms of dynamical variations and wave-convection coupling, we examined longitudinal distributions

of the normalized power spectral density of westward-only-filtered 850-hPa meridional winds averaged in 7.5°S – 2.5°N (Fig. 3a) and the squared coherence between those meridional winds and westward-only-filtered OLR in 10°S – 0° (Fig. 3b). Note that the normalization in Fig. 3a is based on the background spectra calculated by smoothing 40 times with a 1–2–1 filter in frequency. Obviously, we find typical MRG signals around 180° at the 4–5-day periods (e.g., Takayabu and Nitta 1993). Around the SWIO (45° – 60°E), the wind and OLR fields strongly correlate with each other at the 4.5-, 6-, and 8-day cycle (Fig. 3b). In this frequency band, wind fields also show higher power spectra (Fig. 3a). This suggests that convectively

FIG. 3. Longitudinal distributions of (a) the normalized power spectral density of westward-only-filtered meridional winds at 850 hPa in 7.5°S – 2.5°N and (b) the squared coherence between those meridional winds and westward-only-filtered OLR signals in 10°S – 0° . Black squares denote our focused frequency band and region.

FIG. 4. (a) Time–longitude diagram of regressed 850-hPa meridional wind (shading; m s^{-1}) and OLR anomalies (contours; W m^{-2}), based on -2.0 standard deviations of the MRGTD-filtered OLR anomalies at 7.5°S , 50°E . Contour interval is 4 W m^{-2} , with positive (zero) contours dashed (omitted). Meridional wind (OLR) is averaged over 7.5°S – 2.5°N (10°S – 0°). (b) Map of regressed OLR (shading; W m^{-2}), and 850-hPa horizontal wind (vectors; m s^{-1}) and streamfunction anomalies (contours; $\text{m}^2 \text{ s}^{-1}$) on lag 0. Contour interval is $1.2 \times 10^5 \text{ m}^2 \text{ s}^{-1}$, with negative (zero) contours dashed (omitted). (c) Longitude–height section of regressed temperature (shading; K), zonal and vertical wind (vectors; m s^{-1} and Pa s^{-1} for horizontal and vertical components, respectively), and meridional wind anomalies (contours; m s^{-1}) averaged over 7.5°S – 2.5°N on lag 0. Contour interval is 0.35 m s^{-1} , with negative (zero) contours dashed (bolded). The vertical p velocity in vectors is multiplied by 500. Black dashed line indicates the estimated direction of energy dispersion. The presence of shading and vectors in all plots denotes statistical significance at the 95% level.

coupled MRGs and/or TDs with shorter wavelengths than typical MRGs (Kiladis et al. 2009) possibly exist in the SWIO. We also find the spectral peaks of meridional winds at the 5–6-day cycle in 60° – 90°E (Fig. 3a), which indicates that MRGTD dynamics can be realized in the IO. Hence, to isolate MRGTDs in the IO, we used the Lanczos filter spanning westward-propagating periods of 4–10 days in the present study.

We then reconstructed the MRGTD structure in the IO following the method in section 2b. The regressed base point of MRGTD-filtered OLR anomalies was 7.5°S , 50°E , which is in the region with strong coherence between MRGTD dynamics and convection (Fig. 3b). Figure 4a illustrates a time–longitude diagram of regressed 850-hPa meridional wind and OLR anomalies. It is evident that convection is related to southerlies slightly to the east of its convective center, and that the wave packet in wind fields propagates eastward over the IO at about $5^{\circ} \text{ day}^{-1}$. The latter reminds us of the eastward group velocity of MRGs. For these variations, we can roughly estimate the zonal wavenumber of 9–12 and the period of 6–7 days, which reflects the mixed feature of MRGs and TDs in terms of wavenumber–frequency relation.

A horizontal map of convection and 850-hPa circulations for lag 0 (Fig. 4b) suggests that convection is enhanced in the SWIO in conjunction with off-equatorial clockwise circulations and

that cross-equatorial anticlockwise circulations are also established to the east of the SWIO. Note that the horizontal scale of the circulation diminishes around the SWIO, consistent with the wavelength shortening in the WIO found by TSY19. In a zonal–height cross section of equatorial circulations and temperature (Fig. 4c), iso-phase lines of meridional winds incline eastward to 300 hPa in 45° – 60°E . Furthermore, the maximum of meridional wind anomalies along their iso-phase lines directs downward and eastward (black dashed in Fig. 4c), indicating downward-eastward energy dispersion. Notably, a linear theory predicts that a eastward-inclined phase line associated with vertical phase propagation of MRGs corresponds to their downward-eastward group velocity (Holton and Hakim 2012). In terms of the energetics, strong ascent around 300 hPa significantly correlates with positive temperature anomalies for lag 0 (Fig. 4c), implicating efficient baroclinic conversion from eddy available potential energy (EAPE) to eddy kinetic energy (EKE)—that is, strong wave–convection coupling.

To summarize, MRGTDs were actually detected in the SWIO (i.e., the MJO initiation region), so it is reasonable to hypothesize that MRGTDs in the IO can lead to the initiation of MJO events. However, it is not certain whether MRGTDs are active for the initiation process of all the MJO events and

FIG. 5. (a),(b) Maps of 11-day running-mean MRGTD-filtered OLR variance (shading; $\text{W}^2 \text{m}^{-4}$) and OLR anomalies (contours; W m^{-2}) on day -3 for (a) MRGTD-E and (b) MRGTD-NE. Contour interval is 8 W m^{-2} , with positive (zero) contours dashed (omitted). Stippling denotes the statistical significance of OLR variance at the 95% level. The green square in (a) an area used in lagged-regression analysis to capture the representative MRGTD snapshot for MRGTD-E. (c) Time series of the MRGTD-related EKE tendency integrated over 1000–100 hPa in 15°S – 5°N , 50° – 60°E for MRGTD-E (red) and MRGTD-NE (black). Filled markers denote statistical significance from NDJFMA climatology at the 90% level. (d) As in (c), but for the 400–200-hPa EAPE budget terms. $P_e Q_e$ (solid), $P_e P_m$ (solid dotted), and $-P_e K_e$ (broken) are shown. Filled markers for $P_e Q_e$ denote statistical significance from NDJFMA climatology at the 90% level.

what is the nature of the multiscale interaction between MRGTDs and MJO. In the next section, we identify MJO cases initiated in the SWIO via MRGTD–convection coupling and examine how MRGTD dynamics are involved in MJO initiation and propagation over the IO.

4. Impacts of MRGTDs on MJO initiation and propagation

a. Identifying MJO initiation events with active MRGTD–convection coupling

This subsection concerns how many MJO initiation events over the SWIO are related to convectively coupled MRGTDs there. We now focused on the relationship between anomalous intraseasonal convection and MRGTD convective activities in the SWIO. Specifically, to identify MRGTD-related MJO initiation, we imposed a simple condition that OLR anomalies and 11-day running-mean MRGTD-filtered OLR variance anomalies are negative and positive in 10°S – 0° , 45° – 55°E on day -3, respectively, where and when convective signals leading to MJO initiation are evident in general (Fig. 2c). As a result, we could select 26 MRGTD-enhanced cases (hereafter MRGTD-E) and the remaining 21 cases (hereafter MRGTD-NE).

Figures 5a and 5b compare composite OLR anomalies and 11-day running-mean MRGTD-filtered OLR variance anomalies for day -3 between MRGTD-E and MRGTD-NE. As

expected, enhanced MJO convection corresponds well to significant positive MRGTD variance anomalies in the SWIO for MRGTD-E (Fig. 5a), whereas neither convection nor MRGTD activities for MRGTD-NE are outstanding there (Fig. 5b). In addition, the development of MRGTD dynamical fields there before day -3 is confirmed only for MRGTD-E by the positive tendency of 1000–100-hPa MRGTD-related EKE (Fig. 5c), defined as $K' = (\overline{u'^2} + \overline{v'^2})/2$ where primes and overbars denote MRGTD-filtered values and 11-day running means, respectively [the same holds for Eqs. (1) and (2) introduced later]. Hence, MJO convection for MRGTD-E could be triggered by coupling between convection and MRGTD dynamics that spontaneously develop before MJO initiation.

To quantify this MRGTD–convection coupling, we examined the composite time series of MRGTD-related EAPE budget around the SWIO in 400–200 hPa (Fig. 5d), where energy conversion from EAPE to EKE associated with MRGTDs is efficient (Fig. 4c). The budget equation of EAPE ($P_e = RT'^2/2S_p p$) is

$$\frac{\partial P_e}{\partial t} = - \underbrace{\frac{R}{S_p p} \mathbf{V}'_h \cdot \nabla'_h T'}_{P_e P_m} + \underbrace{\frac{R}{p} \omega' T'}_{-P_e K_e} + \underbrace{\frac{R}{C_p S_p p} Q'_1 T'}_{P_e Q_e} + (\text{Resid.})_{P_e} \quad (1)$$

where Q_1 is the apparent heat source computed in a manner of Yanai et al. (1973); C_p is the specific heat at constant pressure;

TABLE 1. Filtering bands for convectively coupled equatorial waves. Positive and negative planetary zonal wavenumbers correspond to eastward and westward phase propagation, respectively. The Lanczos filter is used for the temporal filtering in MRGTDs, and for all other filtering FFT is utilized.

	Planetary zonal wavenumber	Frequency (cycles per day)	Equivalent depth (m)
MRGTD	−1 or less	0.1 to 0.25	—
Kelvin	1 to 15	0.05 to 0.4	8 to 90
ER	−10 to −1	1/96 to 1/10	1 to 90
EIG	0 to 15	0.2 to 0.55	12 to 50
WIG	−15 to −1	0.35 to 0.7	12 to 90

and $S_p \equiv R\bar{T}/(C_p p) - \partial\bar{T}/\partial p$ is the static stability. In Fig. 5d, $P_e Q_e$, the EAPE source through MRGTD diabatic heating, is more active than its climatology after day −10 for MRGTD-E and its feature becomes statistically significant around day −3, which is clearly different from the evolution in MRGTD-NE. Considering that the most of $P_e Q_e$ is converted to EKE via $-P_e K_e$, we can quantitatively conclude that enhanced MRGTD–convection coupling is realized in MJO initiation processes of MRGTD-E cases.

b. Comparison with other equatorial wave activities in MJO initiation and propagation

While MRGTD–convection coupling is outstanding in MJO initiation in the SWIO for MRGTD-E, it still remains uncertain whether MRGTD activities are really more prominent than other convectively coupled equatorial waves in that MJO initiation. For example, Yasunaga (2011) and Dias et al. (2017) suggest that MRG amplitudes are not significantly higher within MJO active convective envelopes than other wave disturbances, although their analyses do not focus on our target region and timing. To examine this question, we computed the variance of OLR filtered for other typical waves [Kelvin wave (Kelvin), $n = 1$ equatorial Rossby wave (ER), $n = 0$ eastward and $n = 1$ westward inertia–gravity wave (EIG and WIG)] with FFT, and compared them with the MRGTD-filtered OLR variance at the initiation phase of the MJO. The temporal running mean was applied to the variance of each wave, and the length of the running mean is 11 days for MRGTD, EIG and WIG, 21 days for Kelvin, and 41 days for ER. The wave filtering bands are summarized in Table 1 (cf. Dias et al. 2017). Note that twice-daily ERA-Interim OLR was utilized to filter EIGs and WIGs.

Figure 6 shows the wave-filtered OLR variances for each wave type, averaged over the area 10°S – 0° , 45° – 55°E for days −5 to −1 (i.e., 5-day running mean on day −3) for 47 detected MJO events, separated into MRGTD-E and MRGTD-NE. The variance of each wave is normalized by its daily climatology corresponding to the above period for each MJO case. For all the MJOs, normalized variances of all wave types are higher than their climatology, but they are not statistically distinct from each other according to the 90% confidence interval based on the 5th and 95th percentiles of the variance distribution calculated from 1000 bootstrapped samples with replacement; the composite of all MJO events do not support

FIG. 6. The wave-filtered OLR variances normalized to daily climatology for each wave type, averaged for days −5 to −1 over 10°S – 0° , 45° – 55°E . Gray, red, and blue bars correspond to all events, MRGTD-E, and MRGTD-NE events, respectively. Confidence intervals are indicated at the 90% significance level.

that MRGTDs are especially more active than other waves. The classification into the two categories partly highlights the difference in the relative activity of MRGTDs to other waves in the SWIO: for MRGTD-E, MRGTDs predominate, even though we still cannot neglect the activity of Kelvin and WIG as a secondary factor.

To discuss the relation between wave dynamics and MJOs for MRGTD-E, we present time–longitude diagrams of composite 1000–100-hPa EKE of each wave and OLR anomalies and EKE tendency in Fig. 7. The length of the running mean in EKE is the same as that used in Fig. 6. Based on Fig. 6, we first examine anomalous EKE fields of MRGTD, Kelvin, and WIG waves around MJO initiation. While positive EKE anomalies of MRGTD and Kelvin waves are observed in the SWIO well before MJO initiation (Figs. 7a,b), those of WIGs are realized at almost the same time as the appearance of negative OLR anomalies (Fig. 7e). This implicates that MRGTD and Kelvin waves could initiate MJO convection externally and that enhanced WIGs may be just response to heating associated with MJO initiation rather than its cause.

Figure 7 also provides an insight into the contributions of wave dynamics to MJO propagation. EKE anomalies of MRGTD, Kelvin, and WIG waves are positive over the whole MJO envelope in the IO, which suggests that those waves may assist MJO propagation. Considering that EKE amplitude is largest for MRGTDs, we can expect that MRGTD dynamics drive MJO convection in the IO more efficiently than the other waves. In addition, the comparison of EKE tendency among the three waves reveals that MRGTDs and WIGs tend to be amplified around MJO convection and that its tendency is most positive for MRGTDs in the IO. These situations raise the idea that an MJO-active phase in the IO appears to have some mechanics that especially develops MRGTDs and thus can be primarily affected by them.

The above arguments somewhat reinforce that it is more worth looking into the MJO–MRGTD interaction in the IO than other equatorial waves for MRGTD-E, although the influences of Kelvin waves should be taken care of secondarily (cf. section 4f). Thus, in the following subsections, of

FIG. 7. Time-longitude diagrams of composite EKE anomalies (thick colored contours; J m^{-2}) and EKE tendency (shading; W m^{-2}) integrated over 1000–100 hPa for each wave and OLR anomalies (thin gray contours; W m^{-2}) for MRGTD-E. (a) MRGTD, (b) Kelvin, (c) ER, (d) EIG, and (e) WIG waves. All values are averaged over 15°S – 5°N . Thick red, black, and blue contours denote positive, zero, and negative EKE anomalies, respectively, and their contour interval is displayed in the bottom of each panel. Contour interval in OLR anomalies is 10 W m^{-2} , with positive (zero) contours dashed (omitted). Stippling denotes statistical significance of EKE tendency from its NDJFMA climatology at the 90% level.

various processes, we especially scrutinize statistics of roles of MRGTDs in the MJO that had tended to be dismissed until TSY19.

c. Temporal relationship between MRGTDs and MJO convection

For the 26 MRGTD-E cases, we examined the relationship between MRGTDs and MJO convective initiation and propagation. We here introduce a new analysis method to grasp the evolution of the MRGTD structure in sections 4c, 4d, and 4f. This is motivated by an idea that simple composite analysis alone, which reveals overall intraseasonal variations, generally has difficulty in capturing the snapshot of MRGTDs embedded in MJO fields at appropriate statistical significance, because MRGTDs have much higher-frequency variations than the time scale at which the overall MJO structure varies and thus the phase relationship between MRGTDs and MJO convection can differ slightly for each MJO case. Hence, to present the representative snapshot of MRGTDs associated with MRGTD–convection coupling in the SWIO, we perform lagged-regression analysis of any MRGTD-filtered variables at each grid point and each lag day for those 26 cases, using 26-sampled MRGTD-filtered OLR anomalies averaged in 7.5° – 2.5°S , 42.5° – 52.5°E (green square in Fig. 5a; area with highly enhanced MRGTD activities) on day -3 (hereafter OLR_{MT}). Note that the regressed values are scaled to -15 W m^{-2} of OLR_{MT} .

Figure 8a shows the time-longitude diagram of composite CWV anomalies and representative 1000–800-hPa meridional wind variations associated with MRGTDs in the equatorial region. In Fig. 8b, the time evolution of composite OLR anomalies (contours) is also presented. Around day -3 , statistically significant anomalous northerlies are realized in 50° – 60°E , which suggests that MJO initiation is likely to be triggered by MRGTD low-level circulations (see Figs. 4a,b). After this MRGTD-related MJO initiation in the SWIO and moistening in the IO by day 0, MRGTD wave packets with significant low-level wind variations successively develop to the east with an eastward group velocity, in accordance with

MJO convective propagation in the IO. In particular, as found by TSY19, the wave packet formation of lower-tropospheric MRGTDs precedes the center of moist anomalies and active convection associated with the MJO. This suggests that low-level MRGTD wind variations could help drive MJO propagation over the IO through convective triggering in a sufficiently moistened environment.

Because several previous works reported that midtropospheric MRGTDs are clearly observed before MJO initiation (Yasunaga et al. 2010, TSY19), we also examined the time-longitude relationship of 600–400-hPa meridional wind variations associated with MRGTDs (Fig. 8b). As the MJO initiation date approaches, midtropospheric MRGTDs gradually amplify around the WIO in conjunction with the slight slowdown of their westward phase speeds (dashed lines in Fig. 8b). These evolutions of MRGTDs are similar to those in an MJO case analyzed by TSY19, although the phase speed change is not as evident here as in that case. As described later, MRGTD modulation is more obviously found in the upper-troposphere rather than in the midtroposphere, which TSY19 also found in an MJO event initiated in November 2011 during the DYNAMO campaign. TSY19 has proposed that this mid-to upper-tropospheric MRGTD modulation is key to the development of low-level MRGTD circulations leading to MJO initiation.

Based on these statistical results, we sought to answer the following questions: 1) What processes and environments develop and maintain MRGTD dynamical fields? 2) Do MRGTDs contribute to the basin-scale moistening in the IO before the start of MJO propagation? 3) How do MRGTD dynamics interact with MJO convective initiation and propagation?

d. Development and maintenance processes of MRGTD dynamical fields

To evaluate the processes that can develop MRGTD dynamical fields before the MJO initiation of MRGTD-E, we conducted MRGTD-related EKE budget analysis with the following equation:

FIG. 8. (a) Time–longitude diagram of composite CWV anomalies (shading; kg m^{-2}) and MRGTD-filtered 1000–800-hPa meridional wind anomalies regressed against OLR_{MT} (contours; m s^{-1}). Solid and broken contours denote 0.75 and -0.75 m s^{-1} , respectively. CWV (meridional wind) is averaged over 15°S – 5°N (7.5°S – 2.5°N). Stippling denotes statistical significance of meridional winds at the 90% level. (b) As in (a), but for regressed MRGTD-filtered 600–400-hPa meridional wind anomalies (shading; m s^{-1}) and composite OLR anomalies (contours; W m^{-2}). Meridional wind (OLR) is averaged over 7.5°S – 2.5°N (15°S – 5°N). Contour interval is 8 W m^{-2} , with positive (zero) contours dashed (omitted).

$$\begin{aligned} \frac{\partial K'}{\partial t} = & - \underbrace{\overline{\mathbf{V}'_h (\mathbf{V}' \cdot \nabla) \mathbf{V}'_h}}_{K_m K_e} - \underbrace{\overline{\mathbf{V} \cdot \nabla K'}}_{A_m K_e} - \underbrace{\overline{\mathbf{V}' \cdot \nabla K'}}_{A_e K_e} \\ & - \underbrace{\overline{\frac{R}{p} \omega' T'}}_{P_e K_e} - \underbrace{\overline{\nabla \cdot (\mathbf{V}' \Phi')}}_{G K_e} + (\text{Resid.})_{K'} \end{aligned} \quad (2)$$

where $\mathbf{V}$ is 3D wind vector; and R is the gas constant. The terms on the right-hand side in Eq. (2) are as follow: $K_m K_e$ is the barotropic conversion from the background to EKE; $A_m K_e$ and $A_e K_e$ are the EKE advection by background and eddy flows, respectively; $P_e K_e$ is the baroclinic conversion from EAPE; $G K_e$ is the EKE dispersion associated with the work done by the pressure gradient force; and $(\text{Resid.})_{K'}$ is a residual including subgrid-scale diffusive processes. The EKE source and sink terms are $K_m K_e$, $P_e K_e$, and $(\text{Resid.})_{K'}$, and all the other terms contribute to redistributing EKE.

Figure 9 shows the domain-mean vertical structure of the EKE budget terms on day -8 when the positive EKE tendency around the SWIO is largest and statistically significant (Fig. 5c). We omit $A_e K_e$ here because its contributions are too small. In raw values (Fig. 9a), we find a relatively large positive net tendency at 600–400 hPa, which is explained almost entirely by $G K_e$. Considering that $G K_e$ is largely positive below

300-hPa level whereas $P_e K_e$ (largely balanced by $P_e Q_e$; not shown) is largest around 300 hPa, it is plausible that upper-tropospheric EKE generated by MRGTD–convection coupling is dispersed downward. Note that this is consistent with the statistical MRGTD characteristics shown in Fig. 4c in terms of the relationship between vertical motions and temperature at 300 hPa and downward-eastward energy dispersion that corresponds to westward-upward phase propagation. As for the latter, the MRGTD structure before the MJO initiation is actually inclined eastward in vertical (as presented in Fig. 11). Last, $K_m K_e$ is responsible for EKE production at 1000–700 hPa.

To understand which processes are more enhanced in the period leading up to the MJO initiation, we determined the deviations of the budget terms from their average for days -30 to -1 (Fig. 9b). Both $P_e K_e$ at 400–200 hPa and $G K_e$ at 900–300 hPa are more active than their 30-day mean, while $K_m K_e$ in the lower troposphere does not deviate far from its 30-day mean. These results suggest the following energetic views: 1) upper-tropospheric EKE production due to positive correlation between diabatic heating and upward motions (i.e., MRGTD–convection coupling) is strengthened before MJO initiation; 2) downward energy dispersion is significantly enhanced in accordance with the process 1, which contributes to the development and maintenance of mid- to low-level

FIG. 9. Vertical profiles of the composite EKE budget terms averaged over 15°S–5°N, 50°–60°E on day –8. The total tendency (dashed black), $A_m K_e$ (green), $K_m K_e$ (blue), $P_e K_e$ (red), $G K_e$ (yellow), and the residual (purple) are shown. (a) Raw values and (b) anomalies from the mean of raw values from days –30 to –1.

MRGTD circulations; and 3) barotropic conversion processes are also important in the lower-troposphere but they are not particularly strengthened before MJO initiation.

Figure 10a shows the horizontal distribution of deviations of $P_e K_e$ and $P_e Q_e$ integrated over 400–200 hPa on day –8 from their 30-day means before day 0. The upper-tropospheric EKE generation via $P_e K_e$ becomes more dominant to the north of the equator around 60°E, which is quantitatively explained by the more production of EAPE from $P_e Q_e$: enhanced MRGTD–convection coupling. In correspondence with the representative MRGTD structure on day –8 in Fig. 10b, which shows MRGTD-filtered 400–200-hPa vertical motions and 1000–800-hPa horizontal wind anomalies regressed against OLR_{MT} , the area with enhanced upper-tropospheric $P_e K_e$ matches the area with upward motions related to lower-tropospheric cross-equatorial southerlies. This relationship between $P_e K_e$ (or $P_e Q_e$) distributions and the MRGTD structure further confirms tight coupling between MRGTD dynamics and convection. In this situation, the distribution of $G K_e$ anomalies at 600–400 and 1000–800 hPa shown in Fig. 10c suggests that mid- and lower-tropospheric $G K_e$ become positive in the equatorial region of 50°–60°E, indicating EKE dispersion responsible for the establishment of MRGTD circulations that can trigger MJO convection.

While the direct cause of the development of MRGTD dynamical fields is enhanced EKE generation and dispersion associated with MRGTD–convection coupling, an environmental factor that could promote such coupling in the WIO before the MJO initiation is still unclear. As suggested by

TSY19, one candidate is a dynamical effect of mid- to upper-tropospheric background zonal winds ($\bar{u}$) on the modulation of the zonally propagating wave structure, based on a linear wave theory (Lighthill 1978) that zonally propagating waves encounter the wave contraction in a background zonal convergence area (i.e., $dk/dx = -k d\bar{u}/dx$ where k is zonal wave-number). This is actually confirmed in Fig. 11, which shows the zonal–vertical sections of equatorial MRGTD meridional winds and 11-day running-mean zonal convergence for days –7 to –5. The zonal wavelength shortening and slower phase propagation are observed in the upper troposphere with zonal convergence, as indicated by the crosses denoting the minima and maxima of meridional winds at 250 hPa. We also find the development of deep structure with upward motions around 60°E on day –6 (Fig. 11b). These findings support the occurrence of MRGTD modulation and continuous wave–convection coupling before MJO initiation, as observed for the transition from MRGs to TDs around the western Pacific (e.g., Takayabu and Nitta 1993; Dunkerton and Baldwin 1995; Sobel and Bretherton 1999; Zhou and Wang 2007). Furthermore, low-level MRGTD northerlies in the IO gradually develop in association with the eastward-inclined vertical wave structure, consistent with robust eastward–downward energy dispersion related to vertical propagation of MRGTDs.

Another candidate that could promote MRGTD–convection coupling is anomalously moistened environments, which appear to be relevant to efficient baroclinic EKE conversion. Figure 12a shows the longitude–height zonal circulation and moisture fields averaged from days –15 to –5. Mid- to lower-tropospheric

FIG. 10. (a) Map of composite 400–200-hPa PeQ_e (shading; $W m^{-2}$) and PeQ_e (contours; $W m^{-2}$) anomalies from their average from days –30 to –1 on day –8. Contour interval is $0.002 W m^{-2}$, with negative (zero) contours dashed (omitted). (b) Map of MRGTD-filtered 400–200-hPa vertical p velocity (shading; $Pa s^{-1}$) and 1000–800-hPa horizontal wind anomalies (vectors; $m s^{-1}$) regressed against OLR_{MT} on day –8. (c) As in (a), but for 600–400-hPa (shading) and 1000–800-hPa GK_e anomalies ($\pm 0.003 W m^{-2}$ with contours). Stippling over shaded areas and vectors in all plots denote statistical significance at the 90% level.

easterly anomalies associated with suppressed convection are intensified in 50° – $100^{\circ}E$, and the moist signals are significantly concentrated in the western IO located at the eastern edge of the African continent.

FIG. 11. Longitude–height sections of MRGTD-filtered meridional wind ($7.5^{\circ}S$ – $2.5^{\circ}N$; shading; $m s^{-1}$) and zonal ($m s^{-1}$) and vertical wind ($Pa s^{-1}$) anomalies (0° – $5^{\circ}N$; vectors) regressed against OLR_{MT} , and composite 11-day running-mean zonal convergence ($10^{\circ}S$ – $10^{\circ}N$; contours; s^{-1}). Contour interval is $4.0 \times 10^{-7} s^{-1}$, with zero contours bolded. The vertical p velocity in vectors is multiplied by 500. Stippling over shaded areas and vectors denote statistical significance at the 90% level. Red and blue crosses indicate the minima and maxima of meridional winds at 250 hPa, respectively.

To investigate this moisture accumulation process, we examine the column-integrated moisture budget for full fields given by

$$\left\langle \frac{\partial q}{\partial t} \right\rangle = -\langle \mathbf{V}_h \cdot \nabla q \rangle - \left\langle \omega \frac{\partial q}{\partial p} \right\rangle - \left\langle \frac{Q_2}{L_v} \right\rangle, \quad (3)$$

where L_v is the latent heat for vaporization, and Q_2 is the so-called apparent moisture sink (Yanai et al. 1973). The angle brackets indicate mass-weighted vertical integration from the surface to 100 hPa. The sum of the last two terms is a “column process” introduced by Chikira (2014). The horizontal advection is further decomposed into zonal and meridional components with three representative time scales as follows:

$$-\langle \mathbf{V}_h \cdot \nabla q \rangle = -\langle [\mathbf{V}_h] \cdot \nabla q \rangle - \langle \mathbf{V}_h' \cdot \nabla q \rangle - \langle \mathbf{V}_h^* \cdot \nabla q \rangle, \quad (4)$$

where the square brackets, prime, and asterisk denote low-frequency, intraseasonal, and high-frequency components, respectively; for any variable $X = [X] + X' + X^*$, X' and X^* are filtered for the periods of 20–100 days and less than 20 days with a 201-point Lanczos filter, respectively, and $[X]$ is defined

FIG. 12. (a) Longitude–height section of composite zonal (m s^{-1}) and vertical wind (Pa s^{-1}) (vectors and shading for vertical p velocity) and specific humidity anomalies (contours; g kg^{-1}) averaged over 10°S – 10°N from days -15 to -5 . Contour interval is 0.1 g kg^{-1} , with negative (zero) contours dashed (omitted). Stippling over shaded areas and vectors denote statistical significance at the 95% level. The vertical p velocity in vectors is multiplied by 500. (b) Time series of the composite CWV budget terms in 10°S – 10°N , 50° – 60°E . The tendency (dashed black), zonal and meridional advection (solid blue and red), and column processes (solid purple) defined as the sum of vertical advection (dotted green) and the residual (dotted brown) are shown. Filled markers indicate statistical significance at the 90% level. (c) Decomposition of column-integrated horizontal advection in (b), averaged from days -17 to -11 , by the low-frequency ($[\mathbf{V}_h]$), intraseasonal ($[\mathbf{V}'_h]$), and high-frequency ($[\mathbf{V}^*_h]$) wind variations. Gray, blue, and red bars represent the total, zonal, and meridional advection, respectively. Confidence intervals are indicated at the 90% significance level.

as $X - X' - X^*$. Note that moisture budget terms were computed with 6-hourly data and then averaged to daily values. Figure 12b shows the time series of moisture budget anomalies averaged over 10°S – 10°N , 50° – 60°E . The large positive tendency is characterized from days -17 to -11 , before MRGTD dynamics begin to be coupled with convection (i.e., after day -10 ; see Figs. 5d and 9–11), and is mostly caused by zonal advection. The time-scale-based decomposition of horizontal advection reveals the dominant contributions of intraseasonal flows (Fig. 12c). This indicates that the anomalous easterlies associated with MJO convective suppression in the eastern IO efficiently collect moisture in the WIO, supporting the enhancement of baroclinic conversion of EKE related to MRGTD–convection coupling in the SWIO. Such a role for intraseasonal easterlies in moistening processes before MJO initiation has been pointed out in previous observational studies (e.g., Zhao et al. 2013; Sobel et al. 2014; Nasuno et al. 2015; Hung and Sui 2018).

In summary, both the MRGTD modulation caused by mid-to upper-tropospheric background zonal convergence in the IO and moisture accumulation in the WIO due to intraseasonal easterly anomalies are involved in enhanced MRGTD–convection coupling. This enhances and maintains low-level MRGTD circulations through efficient downward dispersion of EKE generated by baroclinic conversion in the upper troposphere, which triggers MJO convective initiation over the SWIO. In this context, however, the existence of MRGTDs is already assumed. We argue for possible reasons why MRGTDs tend to be inevitable before MJO initiation by comparing the situation with that of MRGTD-NE in section 5.

e. Basin-scale moistening in the IO before the start of MJO propagation

We now argue if MRGTDs contribute to basin-scale moistening in the IO required for the start of MJO propagation. Figure 13a shows a time series of anomalous column-integrated moisture budget terms in Eq. (3) averaged over the IO. The moistening until just before the MJO initiation (e.g., around day -5) is mostly due to horizontal moisture advection. Time–height section of horizontal advection as well as vertical p -velocity anomalies in Fig. 13b suggests that, prior to the establishment of large-scale deep ascent on day -3 , significant midtropospheric (650 – 450 hPa) moistening via horizontal advection is realized after day -7 . Thus, the premoistening especially in the midtroposphere over the IO results from large-scale shallow circulations, as reported previously (Hagos et al. 2014; Chen and Zhang 2019). The importance of midtropospheric moistening as a favorable condition for deep convection has been pointed out by numerical studies (e.g., Takemi et al. 2004; Kuang and Bretherton 2006) and observations (e.g., Derbyshire et al. 2004; Takayabu et al. 2006).

The spatial patterns of the midtropospheric premoistening and large-scale circulations that are involved in horizontal advection are presented in Fig. 13c. This shows a horizontal map of $\partial q/\partial t$, horizontal moisture advection, and horizontal wind anomalies over the 650 – 450 -hPa layer averaged from days -7 to -3 . Positive $\partial q/\partial t$ anomalies, particularly in the two regions enclosed by blue and red squares are spatially

FIG. 13. (a) As in Fig. 12b, but for the domain mean in 10°S – 10°N , 60° – 90°E and displaying total horizontal advection (solid red). (b) Time–height section of composite horizontal moisture advection (shading; $\text{g kg}^{-1} \text{ day}^{-1}$) and vertical p -velocity (contours; Pa s^{-1}) anomalies averaged over 10°S – 10°N , 60° – 90°E . Contour interval is 0.003 Pa s^{-1} , with positive (zero) contours dashed (bolded). Stippling denotes statistical significance of horizontal advection at the 90% level. (c) Map of composite 650–450-hPa horizontal moisture advection (shading; $\text{kg m}^{-2} \text{ day}^{-1}$), $\partial q / \partial t$ (contours; $\text{kg m}^{-2} \text{ day}^{-1}$), and horizontal wind anomalies (vectors; m s^{-1}) averaged from days -7 to -3 . Contour interval is $0.1 \text{ kg m}^{-2} \text{ day}^{-1}$, with negative (zero) contours dashed (omitted). Stippling over shaded areas and vectors denote statistical significance at the 90% and 95% levels, respectively. (d) As in Fig. 12c, but for 650–450-hPa horizontal advection averaged over 10°S – 10°N , 60° – 90°E from days -7 to -3 . (e) As in (c), but for horizontal advection by high-frequency flows. (f) Decomposition of 650–450-hPa horizontal moisture advection anomalies, averaged over 10° – 5°S , 60° – 80°E from days -7 to -3 , by MRGTD-, Kelvin-, ER-, EIG-, and WIG-filtered wind anomalies.

contributed by horizontal advection, and there are significant easterlies with weak poleward flows. This is reminiscent of the influences of anomalous wind fields related to large-scale suppressed convection around the MC (Zhao et al. 2013; Nasuno et al. 2015; Maloney and Wolding 2015; Takasuka et al. 2018). In fact, further decomposition of horizontal advection in the IO based on Eq. (4) suggests that intraseasonal zonal and meridional winds act primarily on moisture advection (Fig. 13d). Nevertheless, the contribution from high-frequency meridional winds to advection cannot be neglected. In the distribution of horizontal advection by winds with periods less than 20 days (Fig. 13e), we confirm the significant moistening in the red-squared region due to high-frequency wind components.

Because it is expected that MRGTD circulations are responsible for high-frequency meridional winds and the resultant

horizontal advective moistening, Fig. 13f compares anomalous horizontal advection in the red-squared region via wind anomalies filtered for MRGTDs and other equatorial waves. On average, MRGTDs actually explain about a half of moisture advection due to high-frequency disturbances (Figs. 13d,f), although the contribution of MRGTDs is not necessarily distinct from that of other waves in a statistical sense. This result is not contradictory to the finding by TSY19 and Takasuka et al. (2018).

To summarize, the basin-scale premoistening over the IO is evident in the midtroposphere, which is supported by shallow circulations and broadening of moisture via horizontal advection involving midtropospheric flows at two different time scales: anomalous intraseasonal winds associated with suppressed MJO convection around the MC help moisture

advection, and MRGTD circulations convey moisture to the midtroposphere to some extent. These cooperative moistening processes can be key to moisture resurgence leading to the start of MJO propagation.

f. MJO initiation and propagation associated with convective triggering by MRGTD dynamics

After the development of MRGTD dynamics around the SWIO and the sufficient basin-scale moistening in the IO are accomplished, MJO convection starts to propagate eastward over the IO with the aid of lower-tropospheric MRGTD wind variations. Figure 14 shows a southern equatorial (15°S – 0°) time–longitude diagram of composite synoptic OLR anomalies defined as deviations from 11-day running-mean intraseasonal anomalies and 1000–800-hPa MRGTD-related convergence regressed against OLR_{MT} . For the clarity of explanation, representative MRGTD convergence areas and convective cores associated with MJO envelopes are denoted as Conv-(A*, C*, D*) and CC-(A, B, C, D), respectively. Significant westward-propagating Conv-A* in 45° – 60°E before day -3 is followed by the realization of CC-A there, which is recognized as MJO initiation over the SWIO. Afterward, the eastward group propagation of MRGTD convergence is clearly detected in the IO, and it precedes the convective center of the MJO; the CC-C follows the westward intrusion of Conv-C* around 75°E , and the same holds for the relationship between Conv-D* and CC-D around 90°E . These situations reinforce the perspective of TSY19 that the eastward development of low-level MRGTD circulations at their group velocity helps MJO convective propagation in the IO through efficient triggering of convection.

Meanwhile, we should note a role of Kelvin waves in the MJO evolution over the IO. Kelvin-related robust convergence at 1000–800 hPa (green contours in Fig. 14), which is computed by the same lagged-regression analysis for MRGTDs except for the use of Kelvin-filtered anomalies, propagates into the SWIO around day -3 , although Conv-A* is more obvious there before day -3 and is more tightly coupled to CC-A. In addition, a precursor of CC-B is the intrusion of Kelvin waves rather than that of MRGTDs. We thus do not deny the roles of Kelvin waves in the initial push of MJO propagation, even though MRGTD dynamics appear to make a main building block of MJO convection in the IO.

5. Potential factors differentiating MRGTD-enhanced cases and other cases

The statistical analysis in section 4 suggests that the MRGTD-related MJO realization processes revealed by TSY19 are applicable to more than half of all MJO cases initiated in the IO in boreal winter. However, what conditions differentiate MRGTD-E and MRGTD-NE and why MRGTDs are specifically excited before MJO initiation are not clear. These topics are briefly discussed in terms of the difference and similarity in large-scale circulation between the two.

We first examined the intraseasonal meridional circulation over the IO, because the intraseasonal time scale can affect MRGTD activities. Figures 15a and 15b display

FIG. 14. Time–longitude diagram of MRGTD- and Kelvin-filtered 1000–800-hPa horizontal convergence anomalies [shading ($\times 10^{-7} \text{ s}^{-1}$) and green contour with $-3.6 \times 10^{-7} \text{ s}^{-1}$] regressed against OLR_{MT} , and composite OLR deviations from the 11-day running mean (red contours; W m^{-2}). Contour interval in OLR anomalies is 9 W m^{-2} , with positive (zero) contours dashed (omitted). All fields are averaged over 15°S – 0° . Letters A*, C*, D* and A–D indicate MRGTD-related convergence leading to MJO convection and convective maxima building MJO convection, respectively. Stippling denotes statistical significance of MRGTD convergence at the 90% level.

latitude–height cross sections of 20–100-day filtered meridional flow and nonfiltered specific humidity anomalies averaged from days -15 to -5 in MRGTD-E and MRGTD-NE. For MRGTD-E (Fig. 15a), we find the equatorially asymmetric structure characterized as convective suppression in 10° – 5°S and associated cross-equatorial circulations leading to ascent and a moist region in 0° – 5°N . This evident asymmetry can generate MRGTDs through equatorially asymmetric continuous positive heating. For MRGTD-NE (Fig. 15b), although the meridional structure similarly suggests convective suppression in the southern equatorial region, the mid- to low-level circulation in the Southern Hemisphere tends to blow into the more off-equatorial moist region to the south (20° – 10°S). Accordingly, 1000–850-hPa northward flows and associated ascent and moist anomalies in 0° – 5°N are much weaker than in MRGTD-E. Hence, we speculate that the establishment of the clear intraseasonal cross-equatorial circulation effectively activates MRGTD-enhanced MJO initiation processes.

FIG. 15. (left) Latitude–height sections of composite 20–100-day-filtered meridional (m s^{-1}) and vertical wind (Pa s^{-1}) (vectors and shading for vertical p velocity) and nonfiltered specific humidity anomalies (contours; g kg^{-1}) averaged over $60^{\circ}\text{--}90^{\circ}\text{E}$ from days -15 to -5 for (a) MRGTD-E and (b) MRGTD-NE. Contour interval is 0.2 g kg^{-1} , with negative (zero) contours dashed (omitted). Stippling over shaded areas and vectors denote statistical significance at the 95% level. The vertical p velocity in vectors is multiplied by 250. (right) As in (a) and (b), but for maps of 100-day low-pass-filtered 500-hPa vertical p velocity (shading) and 850-hPa horizontal wind anomalies (vectors; m s^{-1}), and the sum of daily climatology and 100-day low-pass-filtered anomalies of 20–100-day-filtered vertical p -velocity variance at 500 hPa (contours; $\text{Pa}^2 \text{ s}^{-2}$) for (c) MRGTD-E, (d) MRGTD-NE, and (e) their difference. Contour interval for (c) and (d) is $8.0 \times 10^{-4} \text{ Pa}^2 \text{ s}^{-2}$ and for (e) is $2.0 \times 10^{-4} \text{ Pa}^2 \text{ s}^{-2}$. Positive (negative) values in (e) are colored with red (blue). Stippling over shaded areas and vectors denote statistical significance at the 90% level.

A plausible factor that could be responsible for the above differences in the intraseasonal circulation and convection is the interannual variability. Figures 15c–e show the horizontal maps of 100-day low-pass-filtered 500-hPa vertical p -velocity and 850-hPa horizontal wind anomalies for MRGTD-E, MRGTD-NE, and their differences. As an indicator of the activity of the intraseasonal circulation, the sum of daily climatology and 100-day low-pass-filtered anomalies of 20–100-day-filtered vertical p -velocity variance at 500 hPa is also plotted. While anomalous interannual circulations are not significant for MRGTD-E (Fig. 15c), more ascent and associated low-level zonal convergence in $15^{\circ}\text{--}5^{\circ}\text{S}$, $60^{\circ}\text{--}90^{\circ}\text{E}$, and more descent in the SWIO, are realized for MRGTD-NE (Fig. 15d). The interannual ascent anomalies in the off-equatorial south-central IO, which is statistically distinct between the two (Fig. 15e), is supportive of the intraseasonal

southward flows into more off-equatorial regions in 1000–500 hPa (Fig. 15b). In fact, the activity of the intraseasonal circulation at $20^{\circ}\text{--}10^{\circ}\text{S}$ is enhanced in places (Fig. 15e), which is almost explained by interannual components rather than the climatology (not shown). This situation in MRGTD-NE is unfavorable for the excitation and equatorial trapping of MRGTDs in the IO.

For both MRGTD-E and MRGTD-NE, a fundamental source of intraseasonal descent anomalies distributed in the Southern Hemisphere is also an interesting topic. As expected, the intraseasonal circulation activity in $60^{\circ}\text{--}90^{\circ}\text{E}$ is basically stronger to the south of the equator than the north in both cases (Figs. 15c,d). This is probably related to the southern-distributed intertropical convergence zone in the IO during boreal winter (e.g., Zhang and Dong 2004); the boreal winter background itself has enough potential to introduce equatorially

FIG. 16. Schematic diagrams of MJO initiation processes for MRGTD-E.

asymmetric convective suppression in an MJO-suppressed phase. After all, whether such anomalous descent to the south of the equator can link with shallow ascent to the north can be affected by the modulation of interannual large-scale circulations. The sensitivity of these environments associated with the interannual variability and seasonality to the excitation of MRGTDs should be further theoretically quantified, which is left for our future work.

6. Summary and concluding discussion

Motivated by a [TSY19](#) case study, which discovered that MRGs can drive MJO convective initiation and propagation in the IO, we examined the generality of the interaction between observed MRGTDs and MJO during the boreal winters in 1982–2012. We first objectively detected 47 MJO cases initiated in the IO using OLR anomalies, with a tracking method similar to that used by [Suematsu and Miura \(2018\)](#) and [Takasuka et al. \(2018\)](#). The lagged-composite analysis suggests that MJO convection tends to be triggered in the SWIO, as found in [TSY19](#). We also statistically confirmed strong MRGTD–convection coupling in the SWIO. These observations are reminiscent of the influence of MRGTDs on MJO initiation and subsequent propagation.

To quantify the contribution of MRGTD-related MJO initiation processes, we classified MJO events into the two categories: MJO initiation with enhanced intraseasonal convection and MRGTD activities in the SWIO (MRGTD-E); and all others (MRGTD-NE). This classification identified 26 MRGTD-E, more than half of all the MJO cases in boreal

winter, in which MRGTD activities were more prominent than other equatorial waves around the IO. Because this suggests that the MRGTD is one possible contributor to the realization of some, if not all, MJO cases, we further analyzed the detailed processes of the MJO–MRGTD interaction for MRGTD-E.

[Figure 16](#) schematically summarizes the MRGTD-related MJO realization processes. Low-level MRGTD circulations, which are realized near 60°E in accordance with the amplification of MRGTDs above the lower troposphere, can trigger MJO convection in the SWIO (bottom in [Fig. 16](#)). The MRGTD-related EKE budget analysis revealed that the baroclinic conversion in the upper-troposphere and geopotential flux convergence in the mid- to lower-troposphere are more enhanced prior to MJO initiation, which suggests that low-level MRGTD circulations develop via downward dispersion of EKE generated by MRGTD–convection coupling in the SWIO (top in [Fig. 16](#)). This strong MRGTD–convection coupling is supported by two elements. One is the modulation of MRGTD characteristics such as the wave contraction, slower phase propagation, and deeper structure associated with background zonal convergence in the upper troposphere, as mentioned in [TSY19](#). The other is moisture accumulation in the western IO due to horizontal advection by easterly anomalies related to large-scale suppressed convection in the eastern IO.

Basin-scale midtropospheric moistening in the IO is also achieved by horizontal advection immediately before the start of MJO propagation, which makes a suitable condition for deep convection. The circulations responsible for the moisture

advection are not only intraseasonal flows associated with large-scale convective suppression to the east of the MJO initiation region (e.g., Zhao et al. 2013; Nasuno et al. 2015; Maloney and Wolding 2015; Takasuka et al. 2018) but also high-frequency meridional winds related to MRGTDs, as reported by Takasuka et al. (2018) and TSY19. Subsequently, low-level MRGTD wind variations with eastward group velocity successively trigger convection to the east. This can explain the propagation of main MJO convective envelopes, at least in the IO (bottom in Fig. 16). Note that Kelvin waves as well play a secondary role in the initial push of MJO propagation.

We also discussed the potential factors of the difference between MRGTD-E and MRGTD-NE. For MRGTD-E, we found the evident cross-equatorial circulation composed of enhanced descent to the south and shallow ascent to the north of the equator on the intraseasonal time scale, which may force MRGTDs through equatorially asymmetric positive heating. That circulation is not as clear in MRGTD-NE as in MRGTD-E; instead, southward flows in the Southern Hemisphere are recognized, which is because of the interannual atmospheric variability: ascent in the eastern IO at 15°–5°S and descent in the SWIO. This circulation can disturb the equatorial trapping of cross-equatorial meridional overturning in the IO, although equatorial asymmetric intraseasonal descent anomalies themselves are found in both MRGTD-E and MRGTD-NE probably due to the boreal winter background.

The main results of this work demonstrate that the MRG-related MJO initiation and propagation reported by TSY19 are true of many other MJO cases. Specifically, on the premise that environmental conditions such as larger-scale circulations than the synoptic scale have a potential for MJO realization (Suematsu and Miura 2018), the MRGTD is one effective factor that determines when and how MJO convection evolves. This finding provides not only helpful information in MJO prediction but also an opportunity to reconsider existing MJO theories. As for MJO initiation, our proposed mechanism needs neither globally circumnavigating Kelvin waves (e.g., Kikuchi and Takayabu 2003; Seo and Kim 2003; Powell and Houze 2015) nor extratropical forcing (e.g., Hsu et al. 1990; Ray and Zhang 2010; Zhao et al. 2013) to trigger MJO convection, consistent with the mechanism denial study of Ma and Kuang (2016). The fact remains, however, that MRGTD-NE cases sometimes do exist, and it is still an open question what processes are key for their MJO convective realization. To completely understand MJO initiation, it is necessary to quantify the comparative roles of several plausible mechanisms by comparing the detailed initiation processes in MRGTD-NE or examining the modulation of MJO characteristics reproduced in long-term numerical sensitivity experiments for realistic situations.

As for MJO propagation, our standpoint supports Pritchard and Yang (2016), which would challenge the moisture mode view, in that an MRGTD-induced propagation mechanism highlights the importance of high-frequency dynamics in triggering convection. This is close to the thinking outlined in Yang and Ingersoll (2013, 2014). It also emphasizes that MJO convection is directly driven by an MRGTD rather than a

large-scale Kelvin wave/response at least in MRGTD-E. Although the moisture mode theory, which implements the Matsuno–Gill response in a dynamical field in advance and explicitly considers only the evolution of moisture or moist static energy (Sobel and Maloney 2013; Adames and Kim 2016), builds on the observed moisture variations (e.g., Kiranmayi and Maloney 2011; Sobel et al. 2014; Yokoi and Sobel 2015), we believe that it is worth modifying that theory to explicitly treat the cross-scale feedback between high-frequency dynamics such as MRGTDs and convective and moisture fields without making assumptions in the dynamical response. To achieve this, it could be helpful to advance the “MJO skeleton” theory (Majda and Stechmann 2009, 2011), which considers the collective effect of convective envelopes associated with synoptic-scale disturbances despite implementation of the longwave approximation. Taking care of dynamical roles of high-frequency waves as well as the Kelvin–Rossby response in MJO convection would lead to a more realistic MJO theory and fruitful insights into the cross-scale interaction and is thus a focus of our future work.

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