Diverse controlling mechanisms and teleconnections of three distinctive MJO types

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Abstract
In this diagnostic study, three distinctive MJO types in boreal winter are documented and their controlling mechanisms and teleconnections are investigated with a synergetic glocal approach. It is revealed that the diverse nature of the MJO primarily results from different tropical-extratropical interactions and associated internal atmospheric processes. Both the type-I and type-II are initiated over the western Indian Ocean (IO) by a dry zone around the eastern IO (EIO), while only the type-I can move out the IO and circulate around the globe. The type-III initiates over the western Pacific (WP) and can circulate the globe. The strong upper-level equatorial westerly wind over the IO-WP, resulting from upstream and extratropical influences, suffocates the type-II MJO within the IO. Whereas, the robust upper-level equatorial easterly wind over the IO-WP, also resulting from upstream and extratropical influences, along with regional convective instability over the WP and the arrival of cold surge over the South China Sea (SCS) foster the development and eastward propagation of the type-III MJO. The downstream and extratropical teleconnections are primarily controlled by the associated convection over the tropical IO-WP sector for the type-I, but also strongly influenced by the conditions over the extratropical WP for the type-II and type-III. Given that the MJO has been traditionally viewed as a tropical mode owing its existence to the coupling between organized convection and large-scale circulations, present findings advocate the MJO as a glocal mode and call for more research on the involved tropical-extratropical interactions in order to better understand and simulate the diverse nature of the MJO.

1 Introduction
Under a rapid changing climate (IPCC 2021), in response to the grand challenges faced by our planet to build a sustainable future, seamless prediction has emerged as a new endeavor for global weather-climate communities (e.g., Hoskins 2013; WMO 2015), which aims to integrate two traditionally separated communities together (i.e., the weather community and climate community). To bridge the gap between the weather forecasting and climate prediction, global weather-climate communities have been spearheading to improve the understanding of intraseasonal variability centered on the Madden–Julian Oscillation (MJO, Madden and Julian 1971, 1972; Lau and Waliser 2012): the most prominent intraseasonal mode in the tropics featuring a slow eastward-propagating multi-scale convective envelope with 30–60-day period.

Since the discovery of the MJO (Madden and Julian 1971, 1972), 50 years’ synergetic community efforts have led to tremendous progresses in the understanding of major candidate processes governing the MJO (e.g., Lau and Waliser 2012; Maloney and Hartmann 2000; Zhang 2005, 2013; Lin et al. 2009; Zhang et al. 2020). However, present-day global weather and climate models still have numerous problems to simulate and forecast the MJO and its global impacts faithfully (e.g., Hung et al. 2013; Ling et al. 2017). The limited progress made in simulating and forecasting the MJO and its global impacts reflects our lack of comprehensive understanding of the nature of the MJO, possibly due to our treatment of the MJO simply as a canonical mode as depicted in Madden–Julian’s seminar papers. Theoretical studies aim to capture the essential features of the MJO (e.g., the dominant spatial pattern, period, and propagating speed as reviewed in Zhang et al. 2020). Most modeling studies intend to reproduce the gross features of the MJO (e.g., variance pattern, power spectrum, and regression
patterns, Hung et al. 2013). In fact, the MJO represents a quasi-oscillating mode with a broad wavenumber-frequency band (e.g., Roundy 2012), manifesting as a phenomenon with diverse event-by-event variants (e.g., Goulet and Duvel 2000; referred as MJO diversity hereafter). An emerging challenge faced by the community is to address the following two essential issues: (1), what are the major processes leading to the diverse nature of the MJO in terms of their onsets, propagations, and teleconnections? and (2), how to faithfully represent these processes in modern weather and climate models? Present study is an effort towards addressing the first issue, focusing on three specific questions: (1), how to categorize different MJO types? (2), what are the key global and local features of individual MJO types? (3), what are the possible controlling factors of each MJO type?

In literature, the onsets of the MJO over the western Indian Ocean (WIO) might have diverse origins: (1), developing locally, or (2), being triggered by tropical circumnavigating or extratropical influences. The MJO may develop locally through so-called discharge-recharge processes involving air-sea fluxes (Shinoda et al. 1998; Li et al. 2008; Fu et al. 2017, 2018b), boundary-layer moisture convergence (Wang and Schlesinger 1999; Kemball-Cook and Weare 2001; Chen and Wang 2019) and radiation-convection feedback (Bladé and Hartmann 1993; Qi and Randall 1994; Maloney et al. 2010; Wang and Li 2020a). A precedent MJO event can also trigger a new MJO over the WIO through its upper-level circumnavigating zonal wind and/or divergence anomalies (Lau and Peng 1987; Hendon 1988; Roundy 2014; Sakaeda and Roundy 2015) and/or low-level Rossby-wave-like response to precedent dry phase around the Maritime Continent (MC, Seo and Kim 2003; Jiang and Li 2005; Zhao et al. 2013). Equatorward-propagating extratropical wave-trains and cold surges were also observed to trigger the onsets of the MJO (Liebmann and Hartmann 1984; Murakami 1988; Hsu et al. 1990; Ray et al. 2009; Gloeckler and Roundy 2013; Ling et al. 2013, 2014; Zhao et al. 2013; Hong et al. 2017).

Once initiated, the MJO propagates eastward, but not all of them can move out of the IO (Rui and Wang 1990; Hendon and Salby 1994; Hagos et al. 2016; Kerns and Chen 2016; Ling et al. 2019). Wang and Rui (1990) first documented the diverse propagating routes of the MJO convection over the MC. Kim et al. (2014) and Feng et al. (2015) found that some MJO events initiated in the IO can’t propagate over the MC, while others can. Hirata et al. (2013) and Fu et al. (2018a) discovered that, no matter in boreal winter or summer, the MJO exhibits three distinctive downstream evolutions on its passages over the MC: (1), smoothly transitioning from the IO to WP (type-I); (2) rapidly decayed over the MC (type-II); and (3) significantly intensified over the MC (type-III). The type-I and type-II MJOs of Hirata et al. (2013) and Fu et al. (2018a) are very similar to the propagating and non-propagating MJOs defined by Kim et al. (2014).

The potential physical processes leading and halting the MJO eastward propagation have been respectively investigated in the contexts of internal atmospheric processes (e.g., Wang 1988; Kemball-Cook and Weare 2001; Hsu and Li 2012; Kim et al. 2014; Feng et al. 2015; DeMott et al. 2018) and large-scale environmental conditions (e.g., Waliser et al. 1999; Matthews 2000; Hirata et al. 2013; Roundy 2014; Chen and Wang 2018; Wang et al. 2019). When focusing on the zonal asymmetry of the disturbance of moisture itself, the boundary-layer zonal moisture asymmetry is deemed as an essential mechanism for the eastward propagating of MJO (Hsu and Li 2012; Hung and Sui 2018; Wang and Li 2020a). While for studies based on the column integrated moisture or moisture static energy (MSE) tendency (Kim et al. 2014; Feng et al. 2015; DeMott et al. 2018; Wang and Li 2020b), the importance of asymmetric zonal moisture in the boundary-layer is ignored. Instead, some other important factors for the eastward propagation have been found, such as, Kim et al. (2014) emphasized the essential role of a robust dry phase around the MC on the propagating MJO. However, Feng et al. (2015) pointed out that the MJO can propagate eastward with or without a robust dry phase. A recent study of Chen and Wang (2018) revealed that a robust frontal walker cell could favor the eastward propagation of the MJO through coupling the IO MJO convection with the strong dry phase ahead.

In order to gain an overall insight about the potential roles of the tropical-extratropical interactions (e.g., Liebmann and Hartmann 1984; Murakami 1988; Hsu et al. 1990; Ray et al. 2009; Gloeckler and Roundy 2013), bottom-up (e.g., Wang 1988; Wang and Schlesinger 1999; Kemball-Cook and Weare 2001; Hsu and Li 2012; Fu et al. 2017, 2018b) and top-down (e.g., Lau and Peng 1987; Hendon 1988; Roundy 2014; Sakaeda and Roundy 2015) influences along with the tropical internal atmospheric processes on shaping the MJO diversity, a synergetic global-to-local (aka, glocal) approach with multiple perspectives (i.e., A large-scale circulation perspective, a tropical-extratropical interaction perspective, a regional perspective, and a moisture-budget perspective) has been taken in this study to comprehensively investigate the diverse MJO features and associated controlling mechanisms and teleconnections. The remaining part of this article is organized as follows. The data and methods used to categorize individual MJO types are given in Sect. 2. The key global and local features, including the associated teleconnections, and possible controlling mechanisms of individual MJO types are detailed in Sect. 3. The last section summarizes our major findings and discusses possible implications and potential pathways forward.
2 Data and methodology

2.1 Data

The datasets used in this study include the daily-mean Outgoing Longwave Radiation (OLR) with 2.5° × 2.5° horizontal resolution (Liebmann and Smith 1996) and the ERA5\(^1\) hourly data with 0.25° × 0.25° horizontal resolution (Hersbach et al. 2018, 2019). The variables from the ERA5 include one-level surface temperature, sea level pressure (Hersbach et al. 2018, 2019). The variables from the ERA5 data are interpolated onto the 2.5° × 2.5° horizontal resolution and accumulated up to daily-mean. All data covers the period from 1985 to 2017.

2.2 Methods

Two steps have been taken to extract the intraseasonal signals: (1), the time mean, and first three harmonics of annual climatological cycle are removed from the raw daily time series; (2), a 20–90-day Lanczos bandpass filter is applied to the preprocessed time series (Duchon 1979).

To examine the teleconnections of the MJO (Sardeshmukh and Hoskins 1988; Hoskins and Ambrizzi 1993; Matthews and Kiladis 1999; Matthews et al. 2004; Li et al. 2009; Hsu et al. 1990; Chang and Chen 1992; Kemball-Cook and Weare 2001; Ray et al. 2009; Ray and Zhang 2010; Ling et al. 2013; Ray and Li 2013; Zhao et al. 2013; Lee et al. 2019), the extratropical Rossby wave trains are traced with the horizontal wave activity flux (WAF) as defined by Takaya and Nakamura (2001). The WAF vectors are nearly parallel to the group velocity of the extratropical Rossby waves, which provide a good way to track the extratropical wave-packet propagation associated with the linear stationary Rossby wave. Previous studies have proven that a large fraction of the extratropical circulation anomalies is a direct response to MJO heating and can be largely explained by the linear dynamics (Borges and Sardeshmukh 1995; Seo and Son 2012). At the same time, the propagation of the nonlinear wave also impacts the initiation and the propagation of the MJO convection (e.g., Wei et al. 2019, 2020), which is not investigated in this study. The upper-level WAF is calculated with the 200-hPa intraseasonal stream function and climatological horizontal winds during the extended winter (NDIFMA).

The boundary-layer moisture budget is assessed with (1000–700-hPa) integrated 20–90-day-filtered moisture tendency equation (Yanai et al. 1973; Hsu and Li 2012). The moisture tendency at a single pressure level is the sum of horizontal and vertical moisture advections and the atmospheric apparent moisture sink \(Q_2\) (Yanai et al. 1973):

\[
\frac{\partial q}{\partial t} = - \nabla \cdot (q \mathbf{v}) - \omega q - \frac{Q_2}{L} \tag{1}
\]

where \(q\) is the specific humidity, \(t\) is the time, \(\mathbf{v}\) is the horizontal wind vector, \(\omega\) is defined as the horizontal gradient, \(p\) is the pressure, \(Q_2\) is the atmospheric apparent moisture sink, and \(L\) is the latent heat of condensation. The column-integration of the [Eq. (1)] derives:

\[
\left[ \frac{\partial q}{\partial t} \right] = - \left[ \frac{\partial q}{\partial x} \right] - \left[ \frac{\partial q}{\partial y} \right] - \left[ \frac{\partial q}{\partial z} \right] - \left[ \frac{Q_2}{L} \right] \tag{2}
\]

where the prime denotes the bandpass-filtered 20–90-day (MJO) perturbations and the square brackets represent the vertical integration in the lower troposphere from the 1000 to 700 hPa. The first two terms of [Eq. (2)] represent the zonal and meridional components of the horizontal moisture advection; \(x\) is longitude; \(y\) is latitude; \(u\) is the zonal wind and \(v\) is the meridional wind. The statistical significances of all composite fields have been assessed with the two-sided Student’s \(t\) test (Brown and Hall 1999).

2.3 The selection of different MJO types

The procedure that we used to select different MJO types is largely adopted from Hirata et al. (2013) and DeMott et al. (2018). First, the 20–90-day bandpass-filtered OLR anomaly is averaged between 10° N and 10° S for boreal winter (NDIFMA). Then, a simple Empirical Orthogonal Function (EOF) analysis is applied to extract two leading modes. They, respectively, represent the MJO cycle with dry peak phase over the IO and the WP (Kessler 2001; Hirata et al. 2013). Third, MJO events are selected and classified into different types based on the principal components (PC1, PC2) of the first two EOFs. For the type-I, a minimum PC1 (day 0 for composite) less than −0.85 standard deviation is followed by a minimum PC2 less than −0.85 standard deviation within 25 days. The type-II only has a minimum PC1 without a minimum follow-up PC2, while the type-III only has a minimum PC2 (day 0 for composite) without a precedent minimum PC1.

For the entire study period (1985–2017), 167 MJO events are selected. Among them, 93 events are classified as type-I, 33 events as type-II, and 41 events as type-III. To ensure the robustness of the events’ classification (e.g., Feng et al. 2015; DeMott et al. 2018), the Hovmoller diagrams of the

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\(^1\) The fifth-generation climate reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF).
intraseasonal OLR anomalies averaged over 10° N–10° S of all 167 selected MJO events are further scrutinized with additional criteria. For the type-I, it is required that the contour of $-4.5 \text{ W m}^{-2}$ should extend from the IO to WP continuously beyond 125° E with a gap less than 5° longitude. For the type-II, the convection should show continuous eastward-propagation in the IO with at least 30° longitude, but not beyond 125° E (Feng et al. 2015; DeMott et al. 2018). For the type-III, the contour of $-4.5 \text{ W m}^{-2}$ starts around the MC and propagates eastward to the WP continuously (Hirata et al. 2013; Fu et al. 2018a). After the second step scrutinization, 114 MJO events are kept with 77, 16, and 21 being the type-I, type-II, and type-III, respectively.\footnote{The details of the step2 case selection are given in Table s1 of the supplementary material. The peak dates and the monthly distribution of all 114 MJO events are given in Table s2 and Table s3 of the supplementary material.}

### 3 Potential factors shape the three MJO types

#### 3.1 Convective features of three MJO types

Figure 1 gives the composite Hovmöller diagrams of OLR anomalies of three MJO types. The type-I (Fig. 1a) is the canonical MJO depicted in many previous studies (Madden and Julian 1972; Lau and Chan 1985; Knutson and Weickmann 1987; Matthews 2000; Inness and Slingo 2003; Hsu and Lee 2005; Wu and Hsu 2009; Sobel et al. 2010; Oh et al. 2012; Hagos et al. 2016; Tan et al. 2020) or the propagating MJO as classified in some recent literatures (Feng et al. 2015; Kim et al. 2014; DeMott et al. 2018). The convection, once initiated, rapidly intensifies over the WP and propagates further east to the central Pacific (CP). Unlike the type-I and type-II, no robust dry zone presents to the east of type-III MJO convection, there is a robust eastward-propagating dry phase that was deemed to play a favorable role for the eastward propagation of MJO convection (Kim et al. 2014; DeMott et al. 2018). For the type-I MJO (Fig. 1b), a relative weak convection initiates in the WIO and shifts eastward, but rapidly decays before reaching 125° E. The type-II MJO resembles the eastward-decaying events of Hirata et al. (2013) and DeMott et al. (2018). There are two unique features of the type-II MJO: (1), a robust precedent event with convection lingering around the dateline along with the type-II MJO convection over the IO\footnote{This feature resembles the composite (Fig. 1b) of Sakaeda and Roundy (2015) with upper-level westerly wind over the western hemisphere.}; (2), a severe dry zone persists over the WP. The dry zone over the WP behaves quite differently for the type-I and type-II, which dissipates rapidly in the former case, but not for the latter. To some degree, the severe dry zone in the type-II may be enhanced by the westward-propagating transient dry disturbances, which were also deemed to halt the MJO convection over the IO to propagate into the WP (e.g., Zhu and Wang 1993; Roundy and Frank 2004; Feng et al. 2015; DeMott et al. 2018). The convection of the type-III MJO (Fig. 1c) initiates over the MC (Kiladis et al. 2014; Hong et al. 2017; Ling et al. 2017) instead of the WIO as in the type-I and type-II. The type-III MJO convection, once initiated, rapidly intensifies over the WP and propagates further east to the central Pacific (CP). Unlike the type-I and type-II, no robust dry zone presents to the east of type-III MJO convection.
3.2 Diverse controlling processes of the three MJO types

3.2.1 A large-scale circulation perspective

Upper-level velocity potential has long been used to represent the large-scale circulations associated with the MJO (e.g., Lorenc 1984; Krishnamurti et al. 1985; Knutson and Weickmann 1987). Figures 2, 3 and 4, respectively, give the composite spatial–temporal life cycles of the 200-hPa velocity potential (hereafter VP200), divergent wind vectors and OLR anomalies for the type-I, type-II, and type-III MJOs. As discovered in Fu et al. (2018a) for three boreal-summer MJO types, the large-scale circulation life cycle of the
type-I MJO (Fig. 2) is alternatively dominated with basin-wide cross-Pacific overturning circulations (e.g., Fig. 2d, g) and regional cross-MC overturning circulations (e.g., Fig. 2b, f, j). The type-II MJO (Fig. 3) is associated mainly with regional cross-MC overturning circulations (e.g., Fig. 3b, e). The type-III MJO (Fig. 4), however, is primarily associated with cross-Pacific basin-wide circulations (e.g., Fig. 4d, f). The onsets of type-I and type-III MJO convection are preceded by upper-level circumnavigating divergence (e.g., Figs. 2d, 4b) (Hendon and Salby 1994; Matthews 2000; and Roundy 2014), but not for the type-II.

3.2.2 A tropical-extratropical interaction perspective

In this subsection, the tropical-extratropical interactions associated with three MJO types are examined in detail with the 200-hPa stream function (hereafter SF200) and associated wave activity flux (hereafter WAF), as given in Figs. 5, 7, 8.

For the type-I MJO at day − 15 and − 10 (Fig. 5a, b), a pair of cyclonic (negative SF200 in NH, positive one in SH) and anticyclonic gyres (positive SF in NH, negative one in SH) straddling on the equator, respectively, develop on the west and east of a robust dry zone over the EIO-MC. These quadrupole cyclonic and anticyclonic gyres consist of the coupled equatorial Kelvin-wave and flanking-Rossby-wave responses to the MJO convection (Knutson and Weickmann 1987; Rui and Wang 1990; Hendon and Salby 1994; Jin and Hoskins 1995; Matthews 2000; Seo and Wang 2010; Seo and Son 2012; Adames and Wallace 2014), leading to robust upper-level equatorial westerly and easterly winds,

\(^4\) NH and SH, respectively, stand for Northern Hemisphere and Southern Hemisphere.
respectively, over the IO and WP (Fig. 6a). A Rossby-wave train is established over northern Pacific-American-Atlantic sector with alternative anticyclonic and cyclonic gyres residing over the tropical WP-CP, extratropical North Pacific, eastern Russia-to-Alaska and eastern North America (Fig. 5a, b). The last three “CAC”^5 gyres over the Pacific-North American (PNA) region features a positive PNA pattern. In the Pacific sector, the WAF vectors indicate that extratropical waves propagate both northeastward to Alaska-North America and equatorward to eastern Pacific (EP). In the Atlantic sector, wave propagation is primarily equatorward. Therefore, the upper-level easterly over the equatorial western hemisphere (Fig. 6a) results not just from eastward-propagating equatorial Kelvin waves, but also the extratropical teleconnections in response to the dry phase over the MC (e.g., Sakaeda and Roundy 2015).

From day − 5 to 0 (Fig. 5c, d), in association with the onset of type-I MJO convection over the IO and the dry zone moving to the WP, a new set of quadrupole gyres with reversed signs develops over the IO-WP sector. At day 0 (Fig. 5d), a pair of cyclonic gyres is established over the tropical WP. The previous anticyclonic pairs have moved along the equator to the western hemisphere and merged with the extratropical anticyclones. The robust equatorial easterly associated with the intensified anticyclones in the African-WIO sector provides a favorable downward forcing to the intensification of the convection over the IO. As the convection over the IO further intensifies and moves to the EIO-MC at day + 5 (Fig. 5e), the quadrupole Kelvin-Rossby-wave response amplifies too. The associated extratropical teleconnections form a “CACA” wave-train over the WNP, north Pacific, north America and north Atlantic. At

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^5 C and A, respectively, stand for Cyclonic and Anticyclonic “gyres”.
For the type-II MJO from day $-15$ to $-10$ (Fig. 7a, b), the upper-level quadrupole Kelvin-Rossby-wave response forms over the tropical IO-WP region associated with the eastward-moving convection over the MC and dry zone over the IO as in the type-I. However, the zonal spans of the anticyclonic pairs over the WP are much smaller than that in the type-I (Figs. 6, 7b). Instead of propagating eastward as in the type-I, the anticyclonic pairs over the WP rapidly disappear in next pentad (day $-5$ in Fig. 7c). During the same period (day $-15$ to $-5$ in Fig. 7a–c), the cyclonic pairs over the western hemisphere along with the equatorial westerly wind (Fig. 6b) rapidly move eastward to enhance the cyclonic pairs over the IO. As the dry zone moves to the MC-WP, the type-II MJO convection initiates over the IO with unfavorable upper-level westerly wind and convergence (Fig. 3d). At day 0 (Fig. 7d), both the convection over the 

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**Fig. 5** Composite horizontal spatial–temporal evolutions of OLR (shading, W m$^{-2}$), 200-hPa stream function (SF200) (contour, CI: $2 \times 10^6$ m$^2$ s$^{-1}$) and the associated wave activity flux (m$^2$ s$^{-2}$) anomalies, from day $-15$ to $+10$, for type-I MJO. The red solid contours represent the positive SF200 (anticyclonic anomaly over NH and cyclonic anomaly over SH), while the blue dash contours represent the negative SF200 (cyclonic anomaly over NH and anticyclonic anomaly over SH). The OLR, SF200 and wave activity flux vectors are shown for those with statistical significance above 5% level. The two boxes mark the locations of two MJO active centers: the red one represents the EIO ($15^\circ$ S–$15^\circ$ N, 70$^\circ$ E–100$^\circ$ E) and the green one represents the MC-WP ($15^\circ$ S–$15^\circ$ N, 120$^\circ$ E–180$^\circ$ E).
Fig. 6 Composite global time-longitude Hovmöller diagrams of 200-hPa zonal wind (shading, m s$^{-1}$) and OLR (contour, CI: 2.5 W m$^{-2}$) anomalies averaged between 10° S and 10° N for type-I (a), type-II (b), and type-III (c), from day $-30$ to $+30$. Only the regions with statistical significance above 5% level are plotted. The dash (solid) lines represent negative (positive) OLR anomalies. The two green vertical lines highlight the IO, and the red one is the 125° E longitude as the threshold to distinguish type-I and type-II MJO events.

Fig. 7 As in Fig. 5, but for type-II MJO.
IO and the dry zone over the MC-WP intensifies, but the upper-level anticyclonic pairs as a Rossby-wave response to the IO convection never manifest as in the type-I. From day −5 to +5 (Fig. 7c–e), a quasi-stationary elongated cyclonic gyre presents over the northern WP-CP as one lobe of a Pacific-Japan-like pattern (Kosaka and Nakamura 2006; Wu et al. 2020) and/or a persistent North Pacific pattern (Higgins and Mo 1997). The induced convection over the subtropical WP may help sustain the dry phase over the equatorial WP through a meridional overturning cell. Near the eastern end of this elongated cyclonic gyre, the enhanced equatorward wave activity maintains a lingering convection over the equatorial CP during the entire life cycle of the type-II MJO convection over the IO (Matthews and Kiladis 1999; Moore et al. 2010). The lingering convection over the CP keeps feeding the IO with the upper-level westerly wind (Fig. 6b) and blocking the dry zone over the WP and the associated upper-level easterly wind moving downstream.

The lack of upper-level easterly “ventilating” flow and the blocking of enhanced dry zone over the MC-WP lead to the rapid decaying of the type-II MJO convection over the IO (Fig. 7e). During the entire life cycle of the type-II MJO, the extratropical downstream response is likely caused by both the quasi-stationary cyclonic gyre over the subtropical WP and the tropical convection over the IO-WP. From day −5 to +5 (Fig. 7c–e), a negative PNA-like pattern with “ACA” wave-train (similar as in Fig. 5e, f of type-I) has presented over the northern Pacific-American-Atlantic sector.

For the type-III MJO from day −15 to −10 (Fig. 8a, b), a weak dry zone over the WP along with a pair of anticyclones straddling over the equatorial EP gradually moves eastward. The anticyclone sitting around 20°N in the African-Indian sector is intensified by the equatorward-propagating wave activity in association with the positive NAO pattern (Fig. 8b, c; Lin et al. 2009), thus establishing an apparent equatorial easterly flow on its southern flank.
(Fig. 6c). At day − 5 (Fig. 8c), the precedent upper-level easterly wind over the equatorial IO is further intensified (Fig. 6c) by the arrival of the eastward-propagating anticyclonic pairs, providing an efficient “ventilating” flow for the rapid onset of type-III MJO convection around the MC. In response to this convection, a cyclonic gyre is supposed to form over the WNP as a Gill-type response (Gill 1980). However, there is an anticyclone over the WNP, which was propagated eastward from eastern China (Fig. 8a–c). At day 0 (Fig. 8d), the convection over the MC-WP amplifies rapidly along with the enhanced anticyclones in the WNP and southern IO. The associated equatorial easterly wind almost covers the entire eastern hemisphere (Fig. 6c). The NH cyclonic response to the convection over the MC-WP is shifted about 40° east of its SH counterpart due to the existence of a quasi-stationary anticyclone over the WNP. Therefore, the upper-level equatorial westerly wind (Fig. 6c) around the dateline is primarily driven by the SH cyclonic gyre. From day + 5 to + 10 (Fig. 8e, f), the convection over the WP rapidly weakens on its way eastward. The associated cyclonic pairs and equatorial westerly flow quickly move into the IO and foster the development of a dry zone. In terms of the NH extratropical teleconnections, the situation is very similar as for the type-II. The downstream impacts are likely caused by both the quasi-stationary anticyclone over the WP-CP and tropical convection of the MJO. During the life cycle of the type-III MJO (Fig. 8c–f), the anticyclone in the northern extratropical Pacific gradually intensifies and moves to the CP. The associated downstream teleconnections over the northern Pacific-American-Atlantic sector also evolve from a negative PNA-like pattern (Fig. 8c, d) to a positive one (Fig. 8e, f).

In order to better understand the possible processes leading to the rapid development of the type-III MJO convection over the MC-WP around the day 0 (Fig. 8d), the associated 850-hPa wind and stream function (hereafter SF850) anomalies from day − 7 to 0 are examined in Fig. 9. At day 0, the anomalous convection and the associated Kelvin-Rossby wave response can be observed over the MC-WP, which resembles the traditional Gill-pattern, but with the pair of Rossby-wave gyres locating slightly eastward (Fig. 9h). At the same time, there is a robust northerly or northeasterly wind anomaly spreading along the East-Asian coasts from the Okhotsk Sea to the SCS. A pair of northeast-southwest-oriented anticyclonic and cyclonic anomalies can be observed over the East Asia and Okhotsk Sea. As shown in Fig. 9a–g, this anomalous anticyclone over East Asia is originated from the Lake Baikal around day − 5, manifested by the positive SF850 and the associated clockwise wind vectors. For the following four days, this anomalous anticyclone is intensified and stretches southward to southern China. The associated strong northerly wind anomaly intrudes deeper into the tropics over the SCS, implying the potential role of East-Asian cold surge on the onset of type-III MJO (e.g., Pang et al. 2018).

To further verify this hypothesis, the composites of the 925-hPa air temperature, horizontal wind, and the sea level pressure (SLP) anomalies are investigated. As shown in Fig. 10, from day − 5 to 0, a positive SLP anomaly originating from east of the Lake Baikal extends southward and reaches to the SCS associated with the amplification of the Siberia High (SH). During the same period, there is a negative SLP anomaly developing in situ over the Okhotsk Sea that indicates the deepening of the Aleutian Low. From day − 3 to 0 (Fig. 10c–f), the cold air intrudes into the southern China and the SCS. The increased SLP and northerly wind and the drop of air temperature over the SCS represent the arrival of cold surge over the SCS. As defined by Chang et al. (2005) and Pang et al. (2018), two cold surge indices in terms of the averaged 925-hPa meridional winds over the longitude of 110° E–117.5° E along 15° N and the averaged 850-hPa meridional wind over the longitude of 110° E–140° E along 15° N are calculated (Fig. 10c). From day − 5 to 0, the increasing amplitudes of these two cold surge indices also testify the arrival of the SCS cold surge. Therefore, this scenario is consistent with previous findings that the SCS cold surge-related northerly wind favors the initiation, intensification, and eastward propagation of the MJO convection over the MC-WP through enhancing the lower-level convergence (e.g., Hsu et al. 1990; Chang and Chen 1992; Shoji et al. 2014; Wang et al. 2018; Pang et al. 2018; Pang and Lu 2019).

### 3.2.3 A regional perspective

In preceding two subsections, the large-scale circulations and tropical-extratropical interactions associated with three MJO types have been examined. In present and next subsections, we will zoom into the active convection region over tropical Indo-western Pacific domain (from 30° E to 150° W) to reveal the regional dynamic and thermodynamic processes shaping the convection initiations and evolutions of three MJO types. In Figs. 10, 11 and 12, the vertical structures of divergence, moisture, equivalent potential temperature (EPT), apparent heat source and moisture sink anomalies (Q1 and Q2, Yanai et al. 1973) along with the regional circulations for three MJO types will be carefully examined.

For the type-I MJO (Figs. 10, 11, 12a–d), the boundary-layer convergence, positive moisture and EPT anomalies appear as early as day −15 in the WIO before the emergence of negative OLR anomaly (Figures not shown). From day −10 to −5 (Fig. 10a, b), this boundary-layer convergence along with the positive moisture and positive EPT (Fig. 11a, b) anomalies extend eastward to the EIO and upward to mid-troposphere. The associated convection
also evolves from shallow and congestus to deep convection (Fig. 12a, b) with an apparent front walker cell to the east of the convection (Chen and Wang 2018). The higher altitude of maximum $Q_1$ than $Q_2$ (Fig. 12b) suggests that eddy vertical flux convergence contributes significantly to apparent heat source, indicating the MJO convection at a rigorous development stage (Johnson et al. 2015). In next two pentads (day 0 and + 5, Fig. 10c, d), the deep convergence zone still resides over the EIO-MC with boundary-layer convergence along with positive moisture and positive EPT (Fig. 11c, d)

Fig. 9 Composite horizontal spatial–temporal evolutions of OLR (shading, $\text{W m}^{-2}$) and 850-hPa wind (vector, $\text{m s}^{-1}$) and 850-hPa stream function (SF850, contour, CI: $2 \times 10^6 \text{ m}^2 \text{ s}^{-1}$) anomalies, respectively, from day − 7 to 0, for type-III MJO. The OLR, SF850 and wind vectors are shown for those statistically significant above 5% level. The two boxes over the warm pool mark the locations of two MJO action centers: the red one represents the EIO (15° S–15° N, 70° E–100° E) and the green one represents the MC-WP (15° S–15° N, 120° E–180° E)
Diverse controlling mechanisms and teleconnections of three distinctive MJO types

anomalies penetrating to the WP (120°E–160°E) already, thus establishing robust rearward-titling structures of the convergence zone, moisture and EPT. A pair of front and backward walker cells develop on the east and west of the convection, respectively. The alignment of maximum $Q_2$ and $Q_1$ indicates that the MJO convection has transitioned from dominant convective to stratiform regime (Fig. 12c, d; Leary and Houze 1980; Johnson and Young 1983; Schumacher and Houze 2003). The systematic positive ($Q_1-Q_2$) in the troposphere also indicates that radiative heating potentially plays an important role on sustaining the MJO convection. During the entire life cycle of type-I MJO convection (Figs. 10, 11, 12a–d), the boundary-layer convergence, positive moisture and convective instability consistently lead the convection (negative OLR anomaly). This result testifies that the boundary-layer convergence (Hsu and Li 2012; Chen and Wang 2018; Wang et al. 2019), positive moisture (Lin and Johnson 1996; Raymond and Torres 1998; Hsu and Li 2012) and convective instability (Kemball-Cook and Weare 2001; Zhao et al. 2013; Wang et al. 2019) likely work in a cooperative way for the development and eastward propagation of the type-I MJO convection.

For the type-II MJO (Figs. 10, 11, 12e–h), weak and scattering boundary-layer convergence and positive moisture anomalies begin to appear over the tropical IO as early as day −10 (red and blue dash lines, Fig. 10e) although there is still robust convergence and divergence anomalies in the upper and lower troposphere, respectively. The boundary-layer convergence is likely forced by the suppressed dry zone around the MC (Zhao et al. 2013). At the same time, weak positive EPT (Fig. 11e) anomalies also appear in the boundary-layer over the IO with robust convective instability (Fig. 12e, the red dash line), but apparent convection has yet to develop. It should be noted that this robust convective
instability is primarily caused by mid-tropospheric drying rather than boundary-layer moistening. In the next pentad (day − 5 in Fig. 10f), the low-level anomalous convergence and moisture intensifies over the IO and slightly extends upward. As shown in Fig. 11f, the vertical structure of the EPT anomalies illustrates that the positive EPT can extends to 300 hPa with its maximum value locating at lower-troposphere. The resemblance between the purple dash and the red dash lines indicates that the positive convective instability is largely resulted from the lower-level convergence and moistening (Fig. 10f). At day 0, fueled by the robust convective instability (Fig. 11f, g), the convection over the IO is further intensified by the positive feedback between the overturning circulation and the diabatic heating. The robust ascending motion along with positive moisture and EPT anomalies occupies the entire troposphere (Figs. 10, 11g).

The MJO convection transitions from dominant convective regime to stratiform regime (Fig. 12f, g). Different from the type-I MJO, the dry zone to the east of the enhanced convection is further intensified (Fig. 11f, g) instead of weakening (Fig. 11b, c). As the backward walker cell disappears, the front walker cell dominates. The enhanced drying and descending motion over the MC-WP hamper the development of leading boundary-layer convergence and moistening (Fig. 10g). At day + 5 (Figs. 10, 11, 12h), the convection over the IO rapidly decays likely due to the lack of leading boundary-layer convergence and moistening as well as the intrusion of severe dry zone over the WP (Feng et al. 2015; DeMott et al. 2018) although there is robust leading convective instability over the MC-WP sector (Fig. 12g, h).

For the type-III MJO (Figs. 10, 11, 12i–l), weak but significant convergence can be observed in the mid-troposphere (700–500 hPa) around 135° E accompanied by a suppressed dry anomaly (Fig. 10i). Robust convective instability develops over a broad region from the WIO to the dateline (Fig. 11l) and is primarily due to the mid-tropospheric drying instead of boundary-layer forcing. At day − 5 (Figs. 10, 11, 12j), the robust tropospheric and
boundary-layer convergence as well as positive moisture over the IO-MC-WP sector can be observed (Fig. 10j). The convective instability dominated by the low-level positive EPT also becomes significant there (Fig. 11j). The strong anomalous convection of type-III MJO establishes rapidly with a broader zonal band (compared with type-I and type-II MJOs). It is worth noticing that for type-II and type-III MJOs, while the positive convective instability dominated by the mid-tropospheric drying anomaly already occurred around day −10 (Fig. 11e, i), the MJO convection did not show up until day −5 when the convective instability is resulted from the positive lower-level EPT (Fig. 11f, j). This might imply that the convective instability induced by boundary-layer moistening drives MJO initiation, which is consistent with the hypothesis that the zonally asymmetric moisture in the boundary-layer may be important to the initiation of the MJO convection (e.g., Zhao et al. 2013).

The convection on the west and east sides of about 135°E are, respectively, dominated by stratiform and convective regimes (Fig. 12j). The robust in-situ convective instability (Fig. 11j), along with the arrival of the SCS cold surge (Figs. 9h, 11k), boosts the rapid intensification of deep convection over the WP at day 0 (Figs. 10, 11k). The corresponding convection also shifts from convective to stratiform regime (Fig. 12k). The persistent positive \((Q_1-Q_2)\) anomalies at MJO peak phases (with dominant stratiform rainfall) of all the three MJO types (Fig. 12c, d, g, k) testify the important role of radiative heating on sustaining the MJO regardless the MJO types. Different from both the type-I and type-II, the type-III MJO has a predominant backward walker cell instead of a dominant front walker cell, likely due to the lack of a robust deep dry zone preceding the type-III MJO convection (Fig. 10j–l). At day +5, the boundary-layer convergence, positive moisture (Fig. 10l) and convective instability (Fig. 11l) over the CP favor the type-III MJO convection to further propagate eastward.
3.2.4 A moisture-budget perspective

The above regional analysis suggests that the MJO convection initiates for type-I at day − 10, type-II and type-III MJOs at day − 5, aligning with the emergence of boundary-layer moistening and lower-level positive EPT associated with convective instability. Moreover, for the type-I and type-III, there are apparent positive boundary-layer anomalies ahead of the eastward-propagating MJO convection, but not for the type-II (Fig. 11). As suggested by many previous studies that positive boundary-layer moisture anomaly or its tendency plays an important role on the initiation (e.g., Kemball-Cook and Weare 2001; Li et al. 2015; Mei et al. 2015; Wei et al. 2019, 2020; Zhao et al. 2013) and the eastward propagation of MJO convection (e.g., Hsu and Li 2012; Wang et al. 2017; Hung and Sui 2018; Chen and Wang 2018; Wang and Li 2020a, b). For different phases or locations of the MJO, the major processes contributing to the boundary-layer moistening varies considerably. Zhao et al. (2013) revealed that the horizontal advection contributes most to the lower-tropospheric moistening which triggers the initiation of the MJO. Whereas the vertical advection term dominates the leading boundary-layer moisture that favors the eastward propagation of MJO convection (e.g., Hsu and Li 2012). The horizontal advection term has a prominent positive contribution around 100° E –180°, indicating its importance for the MJO propagating eastward over the MC and WP (Kiranmayi and Maloney 2011; Kim et al. 2014; Wolding and Maloney 2015; Hung and Sui 2018). More recently, based on a normalized phase evolution analysis, Wang and Li (2020a) argued that for MJO over the EIO, the leading moisture arises from vertical advection, while for the one over WP, it is resulted from surface evaporating process. Therefore, in this subsection, the major processes governing the boundary-layer moisture tendency for three MJO types are investigated through the boundary-layer moisture budget (Figs. 13, 14).

For the type-I (Fig. 13a), the positive boundary-layer moisture tendency leads both the initiation and eastward-propagation of MJO convection. The overall evolution of

![Fig. 13 Composite time-longitude Hovmöller diagrams of OLR anomalies (contour, CI: 2.5 W m⁻²) and 1000–700-hPa integrated moisture budget terms (shading: 1 × 10⁻⁷ g kg⁻¹ s⁻¹) averaged between 10° S and 10° N: a, f, k tendency, b, g, l vertical advection, c, h, m moisture sink, d, i, n horizontal advection, and e, j, o the addition of vertical advection and moisture sink terms for MJO type-I (top row), type-II (middle row), and type-III (bottom row) from day − 30 to + 30. Only the regions with statistical significance above 5% level are plotted. The dash (solid) lines represent negative (positive) OLR anomalies. The two purple vertical lines highlight the IO region, and the red ones are the 125° E longitude as the threshold to distinguish type-I and type-II MJO events.](image-url)
the moisture tendency resembles the horizontal advection term (Fig. 13d), implying the prominent role of the horizontal advection term. Although the magnitudes of the vertical advection and the moisture sink (Fig. 13b, c) are five times larger than other terms, they almost cancel each other for the out-of-phase relationship between them (Fig. 13e) (Hung and Sui 2018; Feng et al. 2015). Moreover, it can be found that during day 0 to +5, the boundary-layer positive moisture anomaly near 110° E has been enhanced (Fig. 10c, d). During this period, only the horizontal advection term

Fig. 14 Composite time-longitude Hovmöller diagrams of OLR anomalies (contour, CI: 2.5 W m$^{-2}$) and 1000–700-hPa integrated moisture budget terms (shading: 1x10$^{-7}$ g kg$^{-1}$ s$^{-1}$) averaged between 10° S and 10° N: a, d, g zonal advection, b, e, h meridional advection, and c, f, i horizontal advection as the addition of the zonal and meridional advection terms for the type-I (top row), type-II (middle row), and type-III (bottom row) from day $-30$ to +30. Only the regions with statistical significance above 5% level are plotted. The dash (solid) lines represent negative (positive) OLR anomalies. The two purple vertical lines highlight the IO region, and the red ones are the 125° E longitude.
tendency east of 105° E (Fig. 13f). The lack of apparent positive i), thus resulting in near-zero boundary-layer moisture tendency east of the IO convection (Zhao et al. 2013; Hung and Sui 2018). On the other hand, the meridional advection leads to the positive advection over the 110° E – 180°, favoring the eastward propagation of the MJO (Kiranmayi and Maloney 2011; Kim et al. 2014; Feng et al. 2015; Wolding and Maloney 2015).

For the type-II (Fig. 13f), the leading positive tendency only exists over the IO region, which is mainly contributed by the horizontal advection (Fig. 13i). The maximum positive tendency around 90° E, at day −10 (Fig. 13f) is also contributed by the moisture sink term although the negative vertical advection plays an opposite role (Fig. 13g, h, j). As in the type-I, the horizontal moisture advection generates positive boundary-layer moisture tendency before the onset of the type-II MJO convection in the IO and to the east of the convection in the WP (Fig. 14i), which is supposed to favor the initiation of convection and its eastward propagation. However, the strong negative vertical advection induced by the dry zone over the WP (Figs. 10f–h, 13g) exceeds the positive moisture sink and horizontal advection (Fig. 13h, i), thus resulting in near-zero boundary-layer moisture tendency east of 105° E (Fig. 13f). The lack of apparent positive boundary-layer moisture tendency east of the IO convection likely hinders the eastward propagation of type-II MJO (Feng et al. 2015). Further decomposition shows that meridional advection plays a larger role than the zonal advection on the positive horizontal advection term (Fig. 14d–f).

For the type-III MJO (Fig. 13k), there is no obvious positive boundary-layer moisture tendency preceding the onset of convection over the MC (from 90° E to 120° E). The positive moisture tendency over the WP-CP may favor the eastward extension of the type-III MJO convection through moistening the boundary-layer (Fig. 10). This result suggests that the boundary-layer process may just play a minor role on the initiation of type-III MJO convection over the MC. It is the upper-level easterly “ventilating” flow (Figs. 6c, 8c, d; Roundy 2014) and the arrival of cold surge-related northeasterly wind trigger the onset of type-III MJO convection over the MC. Similar as the Type-I and Type-II, the positive moisture tendency of the type-III over the 120° E – 170° E is basically caused by horizontal advection (Fig. 13n). The positive tendency east of the dateline is due to the moisture sink (Fig. 13m). Further decomposition (Fig. 14g–i) reveals that zonal advection dominates the positive horizontal advection east of 120° E.

4 Concluding remarks and discussions

In this study, a synergetic glocal approach with multiple perspectives has been adopted to reveal, at the first time, that the diverse nature of the MJO primarily results from the influences of different tropical-extratropical interactions on tropical internal atmospheric processes. Our major findings about the key features (including teleconnections) and controlling factors of three MJO types are summarized into a schematic diagram (Fig. 15).

Following previous studies (Hirata et al. 2013; Feng et al. 2015; Fu et al. 2017; DeMott et al. 2018), a set of combined objective and subjective criteria have been applied to obtain three distinctive MJO types (Table s1 of the supplementary material). Both the type-I (Fig. 1a) and type-II MJO (Fig. 1b) initiate over the western IO (Fig. 15a, b) associated with dry zones on the east. The type-I resembles the canonical MJO depicted in many previous studies (e.g., Madden and Julian 1972; Knutson and Weickmann 1987), which can propagate around the globe. The type-II can’t propagate out of the IO, resembling the non-propagation events in some previous work (e.g., Kim et al. 2014; Feng et al. 2015). In association with the type-II convection over the IO, there is a lingering convection over the CP (Fig. 1b), resembling the composite with precedent upper-level westerly wind over western hemisphere (Sakaeda and Roundy 2015; their Fig. 1b). The type-III (Fig. 1c) initiates rapidly over the WP without any robust dry zone associated (Fig. 15c), which can circulate around the globe.

As illustrated in the schematic diagram of Fig. 15a, the large-scale circulations of the type-I MJO alternatively vary between basin-wide cross-Pacific (e.g., Fig. 2d, g) and regional cross-MC overturning circulations (Fig. 2b, f, j). The extratropical teleconnection and downstream response synchronize well with each other to support the development and eastward propagation of the MJO convection. A robust upper-level easterly anomaly (Figs. 5d, e, 6a), as a downstream and extratropical response to previous dry phase, provides a favorable condition for the onset and amplification of the type-I MJO convection (e.g., Roundy 2014). When the active (suppressed) phase of the MJO moves from the IO to WP, a positive (negative) PNA pattern is established over the northern Pacific-American-Atlantic sector as documented in previous literature (e.g., Hsu 1996; M. and Watanabe 2008; Seo and Son 2012; Tseng et al. 2019; Chen 2021). Under such a favorable large-scale environment (Figs. 2, 5, 6a), the type-I MJO convection (Figs. 10, 11, 12a–d)
develops a robust rearward-titling structure of the convergence zone, moisture and EPT along with a robust front (backward) walker cell on the east (west) of the convection. Very likely, the leading boundary-layer convergence, positive moisture, and convective instability work cooperatively for the development and eastward propagation of the type-I MJO convection. The associated positive boundary-layer moisture tendency (Fig. 13e) consistently leads the onset and eastward propagation of the MJO. The zonal (meridional) advection (Fig. 14a, b) primarily contributes to the onset (eastward propagation).

For the type-II MJO (Fig. 15b), regional cross-MC overturning circulations dominate (Fig. 3). The upstream and extratropical conditions setup a hostile environment for the MJO (Fig. 7). During the onset and development of MJO convection, a robust upper-level westerly flow (Figs. 6b, 7c, d) is established by the associated upstream cyclonic anomalies and a quasi-stationary cyclonic gyre over northern
Asian-WP sector, which suffocates the type-II MJO within the IO (e.g., Sakaeda and Roundy 2015). The extratropical downstream response is largely influenced by the quasi-stationary cyclonic gyre over the extratropical WP with a negative PNA-like pattern (similar as in Fig. 1f of type-I) over the north Pacific-American-Atlantic sector. Under such a hostile large-scale environment (Figs. 3, 7, 6b), the type-II MJO convection (Figs. 10, 11, 12e-h) never develops a robust backward walker cell. A deep front walker cell connects the convection over the IO with the strong dry zone over the WP. Different from the type-I MJO, the drying over the WP is further amplified (Fig. 10f, g) along with the enhanced convection over the IO. The enhanced drying and descending motion over the WP hamper the development of leading boundary-layer convergence and moistening (Fig. 10g) although there is robust leading convective instability (Fig. 11g, h). The convection over the IO rapidly decays likely due to the lack of efficient “ventilating” outflows and leading boundary-layer convergence and moistening as well as the intrusion of deep dry anomaly over the WP (Feng et al. 2015; DeMott et al. 2018). The associated positive moisture tendency in this case (Fig. 13f) only leads the onset of the MJO, not the propagation. Both zonal and meridional advection contribute to the positive moisture tendency (Fig. 14d, e).

The type-III MJO (Fig. 15c), however, is primarily associated with cross-Pacific basin-wide circulations (Fig. 4). Before the onset of the type-III MJO convection, the upstream and extratropical anticyclonic gyres (Fig. 8c, d) establish a robust upper-level easterly flow (Fig. 6c) over the IO-WP sector. The arrival of a strong East-Asian cold-surge (Fig. 9h) along with the favorable upper-level condition boosts the rapid development and eastward propagation of the type-III MJO (Figs. 10, 11, 12i–l). The induced downstream Kelvin-wave-like response triggers the onset of another MJO event (Figs. 1c, 8e, f). During the life cycle of the type-III MJO, the associated NH extratropical teleconnections evolve from a negative PNA-like pattern (Fig. 8c, d) to a positive PNA-like pattern (Fig. 8e, f) over the northern Pacific-American-Atlantic sector, which results from a combined influence of the MJO convection and the quasi-stationary anticyclone over the northern WP-CP. Different from both the type-I and type-II, the type-III MJO has a dominant backward walker cell (Fig. 10j–l) instead of a dominant front walker cell (Fig. 10b–d, f, g), likely due to the lack of a robust deep dry zone preceding the type-III MJO convection. Although no apparent rearward-tilting structure is developed, the boundary-layer convergence, positive moisture (Fig. 10l) and convective instability (Fig. 11l) over the CP lead the type-III MJO convection to propagate further eastward. The associated positive moisture tendency (Fig. 13k) only leads the eastward propagation of the convection, not the onset. The moisture sink and zonal advection are the primary contributor to the leading positive moisture tendency (Figs. 13m, 14a).

Synthesizing the results of Fu et al. (2018a) for three boreal-summer MJO types with the findings of present study, it is noticed that the non-propagating type-II MJO has little to do with the so-called MC barrier effect. During boreal summer, we found that the decaying type-II MJO largely results from the lingering convection over the WNP, which enhances the dry zone over the equatorial WP. The latter hampers the formation of robust leading boundary-layer signals (e.g., convergence, moisture, and convective instability), thus confining the type-II convection within the IO. During boreal winter, it is the upper-layer quasi-stationary cyclonic gyre over the north Asian-Pacific sector and the associated westerly wind suffocates the type-II MJO convection within the IO. The severe dry zone over the WP and lingering convection over the CP inhibits the formation of leading boundary-layer signals (e.g., convergence, moisture, and convective instability), likely resulting in the decay of type-II MJO convection within the IO.

Because the longitudinal range of the MC is around 20° and the geomorphology and topography over the MC are constant for all tropical disturbances, it is unclear why the MC halts some MJO events, but not the others. In fact, the MJO-associated convection can well find alternative channels to move through the MC longitudes (e.g., the marginal seas and SCS in Wang and Rui 1990). The MJO-associated largescale circulations have much broader spatial scales than that of the MC. It is the interplay between the large-scale circulations and the convection, involving the tropical-extratropical interactions, top-down and bottom-up influences, play an essential role for the downstream MJO evolutions. The MC barrier problem most likely is a modeling issue originated from misrepresented model processes.

Given that the MJO owns its existence to the coupling between organized multi-scale convection and large-scale Rossby-Kelvin-wave-like circulations in the tropics (e.g., Wang 1988; Majda and Biello 2004; Adames and Wallace 2014; Roundy 2014) and the MJO is primarily fueled by the moisture from underlying ocean, the tropical atmosphere–ocean interactions play important roles on the initiation, intensity, and propagation of the MJO (e.g., Waliser et al. 1999; Fu and Wang 2004). At the same time, many literatures (e.g., Hsu et al. 1990; Hsu 1996; Straub and Kiladis 2002; Lin et al. 2009; Tromeur and Rossow 2010; Sakaeda and Roundy 2015; Abdillah et al. 2021) have documented that the extratropical circulations can well influence the tropical upper-level and low-level environments. Our synergistic glocal analysis further reveals that the diverse behaviors of the MJO can be largely attributed to different tropical-extratropical interactions. For the type-I, the tropical bottom-up processes dominate the tropical and extratropical large-scale responses with the upper-level tropics and
extra-tropics behaving collaboratively. For the type-II and type-III, on the other hand, the top-down and extratropical influences are essential. In light that the MJO holds the key for the success of subseasonal-to-seasonal and seamless prediction around the globe, our findings highlight the need to expand the traditional view of the MJO as a tropical mode to a glocal mode (e.g., Hsu 1996; Donald et al. 2006) in order to accelerate the understanding and modeling of MJO diverse nature and its glocal influences.

Models are the essential tools used to forecast the MJO and its global impacts. However, it is still unknown to what degree current weather and climate models capture the MJO diversity in nature. Ling et al. (2017) found that latest generation global models failed to reproduce the two favorable onset locations of the MJO in the observations: the western IO and WP, suggesting that present models still have serious problems to reproduce the three MJO types revealed in this study. Xiang et al. (2021) recently revealed that a GFDL model can reproduce some features of MJO diversity and associated teleconnections. As the model skills in forecasting the MJO and its modulations of weather and climate variability largely relies on the model capability in faithfully reproducing the major features of individual MJO events and associated tropical downstream influences and extratropical teleconnections, comprehensive assessments of model simulations of MJO diversity are needed. Multi-model inter-comparisons are recommended to unravel the dominant physical processes misrepresented in current weather and climate models. The pathways to make improvements should be explored with long-term free runs and short-term initialized runs in the context of atmosphere-only and atmosphere–ocean coupled models.

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Data availability All data used in the manuscript can be downloaded from websites. The interpolated daily outgoing longwave radiation (OLR) data was from the NOAA/OAR/ESRL PSD, Boulder, Colorado (Liebmann & Smith, 1996), available at https://www.esrl.noaa.gov/psd/. The ERA5 reanalysis data was from the ECMWF. The ERA5 reanalysis data on single levels is available at https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-single-levels. The ERA5 reanalysis data on pressure levels is available at https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels?tab=overview.

Declarations

Conflict of interest In addition to that stated in the Acknowledgements, the authors declare that no other funds, grants, or other support were received during the preparation of this manuscript; The authors have no conflicts of interests or relevant financial or non-financial interests to disclose.

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