Moistening Processes before the Convective Initiation of Madden–Julian Oscillation Events during the CINDY2011/DYNAMO Period

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ABSTRACT

Convective initiation processes in the Madden–Julian oscillation (MJO) events that occurred during the Cooperative Indian Ocean Experiment on Intraseasonal Variability in the Year 2011 (CINDY2011)/Dynamics of the Madden–Julian Oscillation (DYNAMO) intensive observation period (IOP) were investigated. Two episodes that were initiated in mid-October (MJO1) and mid-November (MJO2) 2011 were analyzed using European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis and satellite data. Moisture budgets in the equatorial Indian Ocean (IO) domain (10°S–10°N, 60°–90°E) were analyzed in detail by separating each variable into basic-state (>80 day), intraseasonal (20–80 day), and high-frequency (<20 day) variations. The quality of the ECMWF reanalysis was also evaluated against the sounding data collected during the field campaign.

In both MJO events, the increase in precipitable water started 8–9 days prior to the convective initiation. Moisture advection decomposition revealed that advection of basic moisture by intraseasonal easterly anomalies and of intraseasonal moisture anomalies by the basic zonal wind were pronounced in these two events. The nonlinear high-frequency terms in the meridional moisture advection were the same order of magnitude as the primary term in the middle troposphere, implying systematic upscale transport of moisture. As a possible mechanism of the acceleration of easterly anomalies, amplification of off-equatorial Rossby wave trains that intruded into the equatorial zone was detected during the preconditioning periods in both MJO events.

1. Introduction

The Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972) is a prominent tropical disturbance that has a broad impact on the global weather and climate (Zhang 2013; Gottschalck et al. 2010). The MJO is related to a wide variety of tropical and extratropical ocean and atmosphere phenomena, ranging from local to global spatial scales and diurnal to interannual time scales. Therefore, it is an important target of extended-range weather forecasting. However, representation of the MJO in numerical models (even state-of-the-art operational models) is not satisfactory (Hung et al. 2013; Zhang et al. 2013). In particular, the accurate simulation and forecast of convective initiation of the MJO is a difficult task (Seo et al. 2009; Gottschalck et al. 2010).

A number of studies have addressed the idea that moisture accumulation is the key to initiation and control of the MJO (Bladé and Hartmann 1993; Kemball-Cook and Weare 2001; Maloney et al. 2010). These have been

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reinforced by observational studies (Kikuchi and Takayabu 2004; Yoneyama et al. 2008; Kiladis et al. 2005). Zhao et al. (2013), analyzing boreal winter MJO cases using the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA), revealed marked increases in lower-tropospheric moisture, temperature, and moist static energy (MSE) over the western Indian Ocean (IO) as a precursor to MJO initiation.

Ling et al. (2013) investigated large-scale precursors that distinguish MJO from non-MJO disturbances using 10-yr Tropical Rainfall Measuring Mission (TRMM) and Interim ERA (ERA-Interim) data. They extracted three significant signals: easterly anomalies in the lower to middle troposphere, surface pressure anomalies with a zonal structure of wavenumber 1, and negative temperature anomalies in the middle to upper troposphere. They referred to these signals as a “dry dynamics mode of the MJO.” The question then arises as to whether these dynamical signals are linked to the moisture buildup. Straub (2013) assessed MJO convective initiation in terms of the Real-time Multivariate MJO (RMM) index (Wheeler and Hendon 2004), which is a standard diagnostic of the MJO, by long-term statistics and case studies. One of the major findings was that “local” processes acted on the convective initiation of the MJO, rather than global dynamics, which are closely monitored by RMM. Alternative methods of distinguishing local convective initiation using OLR or precipitation have been proposed (Gottschalck et al. 2013, hereafter G13; Ling et al. 2013, 2014; Kiladis et al. 2014). One of the most distinctive indicators of the MJO convective initiation is the equatorial concentration of convection (Ling et al. 2013). The effects of local, possibly higher-frequency perturbations on convective initiation need further inspection, along with the large-scale dynamics.

Regarding the origin of MJO initiation, there are two categories of hypotheses: those stressing tropical internal dynamics [e.g., a circumnavigating signal of the previous episode: Lau and Peng (1987); Wang and Li (1994); Kikuchi and Takayabu (2003); Matthews (2008); or local processes in the tropics: Hu and Randall (1994); Jiang and Li (2005); Straub (2013)] and those emphasizing extratropical forcing [e.g., Hsu et al. 1990; Bladé and Hartmann 1993; Matthews and Kiladis 1999; Pan and Li 2007; Lin et al. 2007; Ray and Zhang 2010; Wang et al. 2012]. Ray and Li (2013) evaluated the relative importance of circumnavigating signals and extratropical forcing by performing sensitivity experiments using a general circulation model and concluded that extratropical forcing was more important than were circumnavigating signals in association with mean moisture convergence. They demonstrated that the upper-tropospheric wave energy that originated in the extratropics intruded into the equatorial region to initiate the MJO. However, no detailed description of how it led to convective initiation (e.g., in terms of moisture buildup) was presented. Moreover, extratropical forcing sometimes emerges as specific weather events, such as an outbreak of cold surges (Wang et al. 2012) or in an enhancement of Rossby wave activity (Hsu et al. 1990). Accumulation of case studies is indispensable when tackling the MJO initiation problem.

The Cooperative Indian Ocean Experiment on Intraseasonal Variability in the Year 2011 (CINDY/2011)/Dynamics of the Madden–Julian Oscillation (DYNAMO) field project was conducted to improve our understanding of MJO initiation and its dynamics by combining in situ observational data, objective analysis, and numerical simulations. Improvement of the representation and prediction skill of the MJO in numerical models was another major goal (G13; Yoneyama et al. 2013, hereafter Y13), in which multinational operational centers and research models collaborated. At the Japan Agency for Marine–Earth Science and Technology (JAMSTEC), weeklong forecasts using the Nonhydrostatic Icosahedral Atmospheric Model (NICAM; Satoh et al. 2008, 2014) were made daily (Nasuno 2013), and hindcast experiments were also conducted (Miyakawa et al. 2014). During the intensive observation period (IOP) of the field operations (1 October 2011–15 January 2012), three prominent convective events with intraseasonal time scales occurred. The first and second events have been clearly identified as the MJO (e.g., Y13 and G13).

The objective of this study is to gain insight into the mechanisms of the MJO convective initiation that occurred during the CINDY2011/DYNAMO IOP, with a particular emphasis on the moistening processes leading up to convective initiation. Rich data, including observations and model products for the IOP, allowed for an unprecedented range of materials for detailed analysis. A case study has the advantage of allowing the roles of high-frequency local-scale phenomena in the large-scale dynamics and convective organization to be examined. These tend to be smoothed out in statistical analysis of a large number of samples unless scrupulous analyses are performed. For example, Powell and Houze (2013) revealed that the individual MJO initiation over the CINDY2011/DYNAMO sounding array occurred in less than a week, despite the fact that it appeared to take more than 2 weeks when composited. Understanding similarities and differences in the two events will help isolate the key processes. As a first step, this study focuses on the processes occurring during the preconditioning period over the equatorial IO, namely the large-scale setting in which the MJO events took place. Satellite data and model reanalysis were used for this purpose. The
quality of the reanalysis was also examined using the sounding data obtained from the field observations. The actual triggering of the convective initiation, in which high-frequency and local process are expected to play more critical roles, is not addressed here, but will be discussed in a separate paper using the observation and high-resolution simulation data.

The remainder of this paper is organized as follows. Section 2 documents the data and analysis method used in the present study. Section 3 gives a general description of the target MJO episodes and definition of the convective initiation. Section 4 documents the evolution of the MJO events and moisture budget analysis in the equatorial IO. A discussion and concluding remarks are given in sections 5 and 6, respectively.

2. Data and analysis method

ERA-Interim (Dee et al. 2011) and National Oceanic and Atmospheric Administration (NOAA) daily interpolated outgoing longwave radiation (OLR) data for the period of August 2011–January 2012 were used in the present study. The horizontal resolution of ERA-Interim is 1.0° × 1.0°, and the daily average of the 6-hourly data was utilized. The resolution of the NOAA OLR data is 2.5° × 2.5° daily. The quality-controlled sounding data at Gan Island in the Maldives (Ciesielski et al. 2014a) and TRMM 3B42v7 (Huffman et al. 2007) were also used to evaluate ERA-Interim. The resolution of TRMM 3B42v7 is 0.25° × 0.25° at 3-h intervals, and the daily average was used for the evaluation. Moreover, the 3-hourly globally merged infrared radiation brightness temperature (Tb) dataset with 0.5° × 0.5° resolution, which was created from the half-hourly ~4-km resolution dataset (Janowiak et al. 2001), was used for discussion of high-frequency convective disturbances (section 5).

For the definition of the MJO convective initiation and budget analysis, the daily mean variables were separated into low-frequency (>80 day), intraseasonal (20–80 day), and high-frequency (<20 day) components based on a Lanczos filter (Duchon 1979). Here, the intraseasonal components include both the MJO and non-MJO disturbances with comparable time scale that can affect the MJO (section 4). The moisture budget for the intraseasonal tendencies was diagnosed following Zhao et al. (2013). The moisture tendency equation is

\[ \frac{\partial q}{\partial t} = - \nabla \cdot \mathbf{v} q - \frac{\partial q}{\partial p} - \frac{Q_c}{L} + \text{Diff}, \tag{1} \]

where \( q \), \( p \), \( \mathbf{V} \), \( \omega \), \( \mathbf{V} \), and \( t \) are the specific humidity, pressure, horizontal and vertical velocities, horizontal gradient parameter, and time, respectively. The latent heat of condensation is \( L \), and \( Q_c \) denotes the rate of condensation or evaporation of condensates. Other effects, such as subgrid-scale turbulence, surface fluxes, and any numerical operators are included in Diff. The sum of the last two terms on the right-hand side of Eq. (1) was diagnosed from the advection terms and the net tendency (i.e., the residual term). The net tendency was estimated from the temporal difference in the daily data divided by the time interval. In the tropics, vertical advection and a diabatic heating/moisture sink (residual) are the major terms in balance (Yanai et al. 1973).

Zhao et al. (2013) and Hsu et al. (2011) investigated the interactions between intraseasonal- and synoptic-scale disturbances by a budget analysis of heat, moisture, and kinetic energy by separating each variable into low-frequency (basic state), intraseasonal, and high-frequency (synoptic scale) components and decomposing the advection terms. In a similar manner, zonal, meridional, and vertical advection terms in Eq. (1) were decomposed as

\[ -u \frac{\partial q}{\partial x} = -u \frac{\partial \bar{q}}{\partial x} - u \frac{\partial q}{\partial x} - u \frac{\partial q^*}{\partial x} - \frac{\partial q}{\partial x}, \tag{2} \]

\[ -v \frac{\partial q}{\partial y} = -v \frac{\partial \bar{q}}{\partial y} - u \frac{\partial \bar{q}}{\partial y} - v \frac{\partial q}{\partial y} - v \frac{\partial q^*}{\partial y}, \tag{3} \]

\[ -\omega \frac{\partial q}{\partial p} = -\omega \frac{\partial \bar{q}}{\partial p} - \frac{\partial q}{\partial p} - \omega \frac{\partial q}{\partial p} - \omega \frac{\partial q^*}{\partial p}, \tag{4} \]

where \( u \) and \( v \) indicate zonal and meridional velocities, and \( x \) and \( y \) denote zonal and meridional coordinates. The variable \( X \) is separated into \( X = X + X' + X^* \), with the overbar, prime, and asterisk denoting low-frequency, intraseasonal, and high-frequency components, respectively. All the terms in Eqs. (1)–(4) were separated by frequency filtering. The intraseasonal components of the budget analysis described above are presented in section 4.
3. Two MJO cases

Large-scale conditions during the entire observation period were documented by G13, Y13, and Johnson and Ciesielski (2013, hereafter JC13). The IOP was in a developing La Niña phase and in a neutral Indian Ocean dipole condition, and the SST anomaly was positive over most of the IO (G13, Figs. 1 and 2). The IOP-mean precipitation formed a zonally elongated pattern on both sides of the equator [i.e., a double intertropical convergence zone (ITCZ)–like morphology], with more frequent precipitation in the southern band. A peak was formed around 8°S, 60°E, where the SST gradient was maximized (JC13, Fig. 3). Figure 1 shows time–longitude sections of OLR anomalies and total precipitable water averaged over 10°N–10°S. In (a), anomalies are calculated following Wheeler and Hendon (2004). (c) Time series of the intraseasonal (20–80-day filtered) NOAA OLR (black), precipitable water (green), column-integrated (1000–10 hPa) moist static energy (blue), and U850 (red) in ERA-Interim. Averages in the (10°N–10°S, 60°E–90°E) domain are plotted. The preconditioning period, convective initiation date, and date of minimum OLR anomalies are marked by gray shading, solid lines, and broken lines, respectively (see text for the definitions).
including the CINDY2011/DYNAMO sounding array. Both events were preceded by some convective organization in the upstream region (0°–60°E). The processes over the western IO domain in the MJO convective initiation were investigated by Li et al. (2014, manuscript submitted to J. Climate, hereafter L14).

As for the definition of the MJO convective initiation, Straub (2013) noted that the RMM index amplitude is not necessarily appropriate as an indicator because it primarily reflects dynamical signals. The equatorial concentration of convection is an important measure that can be used to identify the MJO convective initiation. In the CINDY2011/DYNAMO MJO events, a convective regime shift between the ITCZ and MJO was a salient feature (JC13; Y13). In the present study, the date on which the intraseasonal (20–80-day filtered) OLR anomalies averaged over the equatorial IO domain became negative was defined as the initiation date. The interest of this paper is in the beginning of the convective organization that subsequently extended to the MJO scale over the equator. This was around a week before the OLR anomalies reached their minimum values and started migrating eastward. Figure 1c shows a time series of intraseasonal OLR, precipitable water, MSE, and 850-hPa zonal wind averaged in the equatorial Indian Ocean domain (10°S–10°N, 60°–90°E). The MJO convective initiation dates were 18 October and 18 November for MJO1 and MJO2, respectively. The buildup of the MSE prior to MJO convective initiation was evident for both cases. The tendency of the MSE closely followed that of precipitable water, which was supportive of the weak temperature gradient assumption (Sobel and Bretherton 2000). Based on this close relationship, we focused on the moisture tendencies in the following sections. Precipitable water and MSE began to increase after 8 October and 9 November for MJO1 and MJO2, respectively. The periods between these dates and convective initiation (i.e., 9–18 October and 10–18 November) are denoted as “preconditioning” periods in this study. An easterly acceleration of a 850-hPa zonal wind, which was also identified as a precursor distinguishing the MJO from non-MJO events (Ling et al. 2013; Straub 2013), was also detected during the preconditioning periods.

Previous studies have identified systematic biases in the ECMWF operational analyses over the Indian Ocean domain (Nagarajan and Aiyer 2004; Ciesielski et al. 2014b). The average root-mean-square differences were within 0.5 K and 0.5 g kg\(^{-1}\), respectively (Figs. 2a,b). The zonal wind exhibited a mean easterly bias between 950 and 650 hPa with a maximum amplitude of \(1.7 \text{ m s}^{-1}\) at 750 hPa (Fig. 2c). This was similar to that in the ECMWF operational analyses at Colombo, Sri Lanka, for the CINDY2011/DYNAMO period (cf. Fig. 4c of Ciesielski et al. 2014b). Otherwise, the root-mean-square difference of the wind profile was generally less than 1.2 m s\(^{-1}\) throughout the troposphere. Overall, the magnitudes of the biases in ERA-Interim were comparable to or smaller than those reported in the ECMWF operational analyses by previous studies [cf. Figs. 2 and 3 in Nagarajan and Aiyer (2004), Fig. 4 in Ciesielski et al. (2014b)].

To validate the moisture balance in ERA-Interim, the model-predicted surface precipitation rate and that derived from the residual of the moisture budget analysis [i.e., the vertical integral of the third and fourth terms on the right-hand side of Eq. (1)] were compared with TRMM 3B42v7 observations (Fig. 3). The time series of the daily precipitation averaged over the IO domain and the precipitation over the northern/southern sounding array in the analysis generally followed the observed time series. Close examination revealed that the model forecast/diagnosis tended to overestimate (underestimate) the precipitation during the suppressed to preconditioning (active) period of the MJO (Fig. 3). The weaker contrast between the former and latter periods was presumably related to the convective representation in the model. Another possibility would be that the storage of condensates in the former period and the evaporation of the loading condensates in the latter period in the real atmosphere (McNab and Betts 1978; Johnson 1980; Johnson et al. 2014) were not correctly represented in the model. Meanwhile, Xu and Rutledge (2014) showed that the TRMM products underestimated (overestimated) rainfall during the convectively suppressed (active) period in comparison with the shipborne rainfall data from the Research Vessel Roger Revelle (their Fig. 4), presumably due to the insufficient detection of shallow clouds (high cloudiness of upper-level clouds) in the microwave and infrared data. This tendency in TRMM products may partly elucidate the difference between ERA-Interim and TRMM 3B42v7. Based on these results, it can be concluded that the use of ERA-Interim for moisture budget analysis does not pose a critical problem.
4. Moisture budget analysis

a. Evolution over the IO

Time–height sections of intraseasonal $\omega$, $q$, and temperature $T$ are presented in Fig. 4. Subsidence and negative moisture anomalies at the beginning of the preconditioning periods (Figs. 4a,b) transitioned into upward motion with positive moisture anomalies 2–3 days before the date of convective initiation. In the boundary layer (950–1000 hPa), upward motion and warm anomalies started during the early preconditioning stage (Figs. 4a,c). These precursor signals were consistent with those in the climatological MJO (Zhao et al. 2013), but with a slower evolution of moist anomalies and the existence of the stable layer (i.e., positive vertical gradient of $T$) at 650–850 hPa, especially in MJO1. These imply more convectively suppressed conditions prior to these MJO episodes than normally occur. After the convective initiation, upward motion in the upper troposphere collocated with positive temperature anomalies (Figs. 4a,c), suggesting a positive feedback between upward motion and latent heat release.

An intriguing feature of Fig. 4b is that the height of the peak intraseasonal moisture anomalies occurred in the middle troposphere (750–600 hPa). Figure 5 presents the frequency–altitude distributions of the relative humidity (RH) in the sounding data at Gan Island. The lower troposphere (1000–800 hPa) was characterized by frequent occurrences of 70%–90% RH regardless of the MJO life cycle (Fig. 5). The frequency in the middle troposphere (800–500 hPa) ranged from 30% to 90% RH in the suppressed phase (Fig. 5, top), which shifted to 60%–100% RH in the preconditioning period (Fig. 5, middle), and to frequent occurrences of $>90\%$ RH in the active period (Fig. 5, bottom) for both MJO events. Thus, the variations in moisture in the middle troposphere were distinguishable in the unfiltered field observation data. Shifts in the range of RH according to the MJO life cycle were also significant in the upper troposphere (Fig. 5). Because specific humidity generally decreases with height, the peak intraseasonal anomalies occurred in the middle troposphere (Fig. 4b). A question then arises as to what was the major source of the middle-level moistening, particularly, during the preconditioning period.

The moisture budgets over the target IO domain are presented in Fig. 6. The evolution of vertical advection corresponded closely to the evolution of vertical motion (Figs. 4a and 6a), with positive vertical advection reaching the upper troposphere by the convective initiation date. In the preconditioning period, moistening due to horizontal advection was of the same order of magnitude as the net tendency $\partial q/\partial t$ in the middle...
troposphere (Figs. 6b,c). Thus, horizontal advection acted as a major source of middle-level moistening in the preconditioning period of the two MJO episodes when the drying tendency due to vertical advection was still pronounced in the free atmosphere (Figs. 6a,b). After the convective initiation, the vertical component became the major term of moistening, significantly exceeding the net tendency (Figs. 6a,c), and was nearly in balance with moisture consumption by condensation, or the residual term of the moisture tendency equation [Fig. 6d; intraseasonal component of the third and fourth terms on the rhs of Eq. (1)], as argued by Yanai et al. (1973).

The zonal and meridional moisture advection terms are separately plotted in Fig. 7. Zonal advection caused deep moistening throughout the preconditioning period, which started from the middle troposphere (around 600 hPa; Fig. 7a). Meridional advection also contributed to moistening in the free atmosphere within several days before and after the convective initiation date, whereas drying effects were pronounced below 850 hPa (Fig. 7b). Specific to MJO1, the meridional advection was enhanced in the lower troposphere from early October. This feature was absent in MJO2. These differences between the two MJO episodes are discussed again later in this section. After the mature stage (defined by the minimum OLR; Fig. 1c), horizontal advection became negative (Figs. 6b and 7), representing the zonal divergence in the lower troposphere and intrusion of dry air from the off-equatorial regions.

Further analysis was made with a decomposition of moisture advection by frequency filtering (section 2; cf. Zhao et al. 2013). Figure 8 shows the vertical integrals of the intraseasonal components of the decomposed terms [rhs of Eqs. (2)–(4)] in the major layer of positive horizontal advection (425–787.5 hPa; Figs. 6b and 7) and in
the layer below (1000–787.5 hPa). The results were insensitive to slight modifications of the vertical range. Hereafter, they are denoted as the “middle” and “lower” troposphere, respectively. The primary component in the middle troposphere was the transport of basic moisture by the intraseasonal wind [the fourth terms on the rhs of Eqs. (2)–(4); Fig. 8, left panels]. In these preconditioning periods, vertical transport was negative (nearly zero or slightly positive) in the middle (lower) troposphere (Fig. 8, bottom panels) and was canceled out by zonal and meridional terms (Fig. 8, upper and middle panels, respectively). In the zonal component, the transport of intraseasonal moisture by the basic wind $[-\pi(\partial q/\partial x)]'$ [the second term on the rhs of Eq. (2)] amounted to half of the primary component $[-u'(\partial q/\partial x)]'$ in the middle troposphere. As for high-frequency effects, it was noteworthy that the transport of high-frequency moisture by high-frequency winds [the ninth term on the rhs of Eqs. (2)–(4)] was evident, especially for the meridional component in the middle troposphere, $[-v'q/\partial y)]'$. Advection of high-frequency moisture by the intraseasonal wind $[-u'(\partial q/\partial x)]'$, $[-v'(\partial q/\partial y)]'$ was also positive (amounting to approximately 30% of the fourth

**Fig. 4.** Time–height section of the intraseasonal (a) omega, (b) specific humidity, and (c) temperature in ERA-Interim averaged over the (10°N–10°S, 60°–90°E) domain. The solid and broken lines indicate the convective initiation date and the beginning of the preconditioning period, respectively.
terms) in the middle troposphere (Fig. 8, top-left and middle panels). This suggests nonnegligible effects of synoptic-scale and mesoscale convective disturbances on the moistening processes that precondition the manifestation of the MJO-scale convective envelope. Meanwhile, salient features specific to each MJO event were also detected, such as $\left[-\nu\left(\frac{\partial q}{\partial y}\right)\right]'$ in MJO1 (Fig. 8a, middle-left panel) and $\left[-u^*\left(\frac{\partial q^*}{\partial x}\right)\right]'$ in MJO2 (Fig. 8b, top panels). In the following, further analysis of the major advection processes is presented.
b. MJO1

The leading terms of horizontal moisture advection during the preconditioning period (i.e., $-\overline{u'(\partial q' / \partial x)'}$, $-\overline{u'(\partial q / \partial x)'}$) were associated with low-frequency and intraseasonal variations in the middle troposphere. Figure 9 shows horizontal distributions of intraseasonal moisture anomalies ($q'$) and low-frequency winds ($\overline{u, v}$) at 600 hPa (the peak level of horizontal advection), together with intraseasonal OLR anomalies. In early October, moist anomalies corresponding to the southern ITCZ existed along 7.5°S, where a basic easterly prevailed (Fig. 9a).
The moist anomalies extended westward (30°–60°E), where clouds were formed (Figs. 9e,f). These clouds did not develop into a persistent robust MJO-scale convection, but dissipated there leaving moist anomalies in the free atmosphere. This loose convective organization continued over the western IO (Fig. 1a). The moist anomalies were transported eastward along the westerly axis (0°–5°N in this season), reaching the target IO domain in middle October (Figs. 9b,c). A large-scale convective envelope was established on 18 October (Fig. 6g). The moisture advection \( -\overline{u} (\partial q/\partial x) \) averaged during the preconditioning period was most evident in the northwestern part of the target domain (Fig. 9d).

The middle-tropospheric basic moisture \( q \) and intraseasonal wind anomalies \( (u', v') \) during the preconditioning period are presented in Fig. 10a. A moisture maximum was located over the Maritime Continent and gradually decreased westward over the IO. A moist tongue associated with the southern ITCZ was formed in the southern part of the equatorial IO (0°–10°S, 60°–90°E), which is characteristic of this season (JC13; Y13). The equatorial western IO was covered with easterly anomalies (40°–70°E; Fig. 10a). Horizontal plots of the advection terms are presented in Figs. 10b–d. The zonal advection term \( -u' (\partial q/\partial x) \) was pronounced in the western part of the target IO region (Fig. 10b). Thus, the two major zonal advection terms (Figs. 9d and 10b) account for the moistening peak around 60°–70°E (Fig. 10e). Meanwhile, \( -v' (\partial q^*/\partial y) \) (Fig. 10c) and \( -v' (\partial q/\partial y) \) (Fig. 10d) were responsible for the moistening peak along 5°S around 80°–90°E (Fig. 10c): the former suggests that cloud formation and nonlinear moisture transport associated with high-frequency disturbances in this local domain contributed to the MJO-scale moistening during the preconditioning period, while the latter is associated with westward and southward transport of \( q \) by intraseasonal winds. The equatorial easterly anomalies were connected to local circulations in the Southern Hemisphere (e.g., 45°–65°E, 100°–120°E in Fig. 10a), which is a marked characteristic of MJO1. This issue is discussed again in section 5.
c. MJO2

The horizontal distributions of intraseasonal moisture anomalies ($q'$) and low-frequency winds ($\vec{u}, \vec{v}$) at 600 hPa in the preconditioning period of MJO2 are presented in Fig. 11, in a similar manner as those in MJO1 (Fig. 9). A positive moisture anomaly corresponding to the southern ITCZ existed on 12 November (Fig. 11a). Moisture was transported westward along 5°–10°S, and clouds were loosely organized at 30°–60°E (Figs. 11e,f and 1a). Moist anomalies extended eastward over the target IO domain within a few days (Figs. 11b,c), leading to the establishment of MJO-scale convection (negative OLR anomalies) by 18 November (Fig. 11g), although convection was not yet fully concentrated on the equator, but was still active in the ITCZ. These processes were similar to those in MJO1, but with a more distinct contrast between convective and suppressed phases and

![Fig. 9](image-url). (a)–(c) Basic (>80 day) wind (vector) and intraseasonal (20–80 day) moisture at 600 hPa from ERA-Interim and at 6-day intervals (6–18 Oct 2011). (d) Moisture advection $[-\overline{\vec{u}(\partial q/\partial x)}]$ averaged during 9–18 Oct 2011. (e)–(g) As in (a)–(c), but for intraseasonal NOAA OLR. Squares indicate the target IO domain.
a more rapid transition in MJO2. It is also noteworthy that the southward shift of the westerly axis, which was associated with the complete withdrawal of the boreal summer monsoon, allowed more equatorially concentrated eastward transport of moisture (Fig. 11d) in November than in October.

Figure 12a shows the middle-tropospheric $\tau$ and $u'$, $v'$ fields in the preconditioning period of MJO2. Compared with the basic fields of MJO1 (Fig. 10a), rich moisture was concentrated along the equatorial zone (Fig. 12a), as well as basic westerlies (Fig. 11), and wind fields were greater in zonal scale and in magnitude in the preconditioning period of MJO2. The equatorial easterly anomalies were connected to off-equatorial large-scale gyres on both sides of the equator (Fig. 12a). Again, the two major zonal advection terms (Figs. 11d and 12b) accounted for the moistening peak around 60°–70°E (Fig. 12c), and $[-\nu'(<\delta q/\delta y>)']$ (Fig. 12c) was pronounced in the equatorial 70°–85°E region. Along the equator, zonal divergence and meridional convergence centered around 90°E were evident (Fig. 12a), and $[-\nu'(<\delta q/\delta y>)']$ was generally negative in the target IO region (Fig. 8b middle-left panel). This elucidates the difference from MJO1 (Fig. 8a, middle-left panel, and Fig. 10d).

Figures 13 and 14 compare the lower-tropospheric basic moisture and intraseasonal wind. In MJO1, easterly anomalies were less apparent than those in the middle troposphere, with meandering flow along the equator (Figs. 10 and 13). The flow patterns in MJO2...
were similar at both levels (Figs. 12 and 14). These are consistent with more vertically coherent structure in MJO2 than in MJO1 (Figs. 4, 6, and 7). The larger acceleration rate of upward motion (Fig. 4a) allowed the more rapid buildup of moisture in MJO2 than in MJO1 (Figs. 6c, 9, and 11). Such differences between the two events are attributable to the more suppressed thermal conditions in MJO1 than in MJO2 (Fig. 6), as well as the seasonal march (e.g., meridional distributions of $\bar{q}$ and $\bar{u}$) and the activity of equatorial and off-equatorial disturbances. Superposition of multiple factors may have been required for the outbreak of intraseasonal-scale convection for MJO1.

5. Discussion

a. Origin of intraseasonal easterly anomalies

In both MJO1 and MJO2, equatorial easterlies were enhanced around a week before the convective initiation, and contributed to middle-tropospheric moistening.
on intraseasonal time scales. The easterly anomaly in the lower and middle troposphere preceding MJO convective initiation by 20 days is known to be a precursor of primary MJO episodes (Ling et al. 2013). Furthermore, Straub (2013) noted that the lower-tropospheric easterly anomalies appear as a “local” signal leading to MJO initiation over the IO.

Figure 15 shows time–longitude sections of intraseasonal zonal velocity at different levels. In the upper troposphere, a large-scale eastward-propagating signal was apparent (Fig. 15a). This indicated the existence of circumnavigating dynamical disturbances at this level. However, the upper-tropospheric convergence prevailed over the IO in the preconditioning period of MJO1 and MJO2, which was not favorable to convective organization (L14). In the lower to middle troposphere, however, the local enhancement of slowly eastward-propagating signals at 50°–120°E was more pronounced than was the global circumnavigation (Figs. 15b,c). In particular, eastward propagation was not even clear over the IO for MJO1. These results suggest that the easterly anomalies were not solely due to a first baroclinic circumnavigating dynamical disturbance, but were affected by local processes in the middle to lower troposphere. Zhao et al. (2013) argued that a Rossby wave response to negative convective heating over the eastern IO induced equatorial lower-tropospheric easterly anomalies over the western IO. In fact, positive OLR anomalies over the central IO were clear in the suppressed period prior to MJO2 (Fig. 11e) with large-scale gyres (Fig. 14), which is consistent with this scenario (L14).

As another possible mechanism of the easterly accelerations, modulation in the off-equatorial wave activity...
is proposed here. Figure 16 shows a sequence of intraseasonal height and wind fields during the preconditioning period of MJO2 at 600 hPa. The height anomalies formed synoptic-scale wave trains along 10°–30°N, accompanied with geostrophic wind anomalies on 9 November (Fig. 16a). The positive height anomaly centered at 25°N, 50°E developed during the preconditioning period (Figs. 16b–d). Accordingly, the easterly anomalies along the southern edge of the height anomaly reaching the equator remained intense. After the MJO initiation date (Figs. 16d,e), positive height anomalies merged together over 40°–140°E, and the equatorial easterly anomalies extended eastward. These accounted for the behavior of the equatorial zonal winds in the time–longitude plot (Fig. 15b). The wave amplification occurred in less than a week as an episodic event and was more detectable in the middle troposphere than in the upper and lower troposphere. Its role in MJO convective initiation is worth closer further consideration.

The situation in MJO1 showed several different aspects. Figures 17 and 18 present the intraseasonal 600-hPa meridional winds and a sequence of OLR and wind anomalies during the preconditioning period, respectively. The equatorial easterlies appeared at 30°–70°E, as a part of synoptic-scale wave trains on either side of the equator (Fig. 17). In the Southern Hemisphere, the wave train traveled from Madagascar to the southwest of Sumatra. The circulation over Madagascar matched the low OLR signal in the preconditioning period closely (Figs. 18c–e). It can be interpreted as a dynamical response to the convective organization that started in early October (Figs. 18a,b). Off-equatorial convection was diminished as the equatorial MJO convective envelope developed (Fig. 18f). Thus, off-equatorial convective organization and its wave response possibly favored MJO convective initiation for MJO1 through easterly accelerations in the middle

![Figure 13](image1.png)  
**FIG. 13.** As in Fig. 10a, but for 850 hPa.

![Figure 14](image2.png)  
**FIG. 14.** As in Fig. 13, but for 10–18 Nov 2011.

![Figure 15](image3.png)  
**FIG. 15.** Time–longitude sections of the intraseasonal zonal wind at (a) 200, (b) 600, and (c) 850 hPa from ERA-Interim. The averages at 10°N–10°S are plotted.

![Figure 16](image4.png)
troposphere. The cloud organization around Madagascar was not extraordinary in boreal winter, as was seen in the climatological OLR pattern in phase 8 of the MJO (Wheeler and Hendon 2004; Zhao et al. 2013). In pattern and horizontal scale, the corresponding wave train was also similar to those that robustly appear in boreal winter (e.g., Fukutomi and Yasunari 2013). However, the periods of the observed wave trains had a broad range, and classification of the present case is difficult.

Previous studies based on composite analysis (Zhao et al. 2013) and numerical simulations (Ray and Zhang 2010) discussed the roles of extratropical forcing in the MJO convective initiation in terms of barotropic energy conversion. These studies demonstrated that the upper-tropospheric wave energy intruded into the equatorial region and transferred to the intraseasonal perturbations there. Meanwhile, case studies of individual MJO initiation (Hsu et al. 1990; Wang et al. 2012) showed a chain of extratropical and tropical processes, such as episodic amplification of the upper-tropospheric Rossby wave train or an outbreak of cold surge events over the Indo-Pacific domain. The present study showed that the extratropical disturbances possibly contributed to MJO convective initiation through local moistening in the middle troposphere. The Rossby wave downstream forcing scenario may also apply to MJO2, especially in the lower troposphere (L14).

b. Roles of high-frequency disturbances

The budget analysis over the IO revealed that the intraseasonal vertical advection associated with high-frequency variations \(-\omega(\partial q^*/\partial p)^t\), \(-\omega^a(\partial q^*/\partial p)^t\) were insignificant compared with the primary term \(-\omega(\partial q^*/\partial p)^t\), whereas in the horizontal components, nonlinear advection of high-frequency moisture \(-\left(V^t + V^*\right) \cdot Vq^t\) was positive in the middle troposphere (Fig. 8). Investigations of high-frequency disturbances utilizing the CINDY2011/DYNAMO field observation data have been undertaken (e.g., JC13; Zuluaga and Houze 2013; DePasquale et al. 2014). Zuluaga and Houze (2013) demonstrated that westward-moving convective episodes with a 2–4-day period appeared repeatedly over the observation site. JC13 reported that westward-moving 2-day disturbances were pronounced in MJO1, whereas 4–5-day disturbances were more frequent in MJO2. By analyzing radar data, DePasquale et al. (2014) discussed the roles of convectively coupled Kelvin waves in deep moistening during the developing phase of the MJO events. These high-frequency disturbances may be responsible for the moisture transport detected in this study (Fig. 8), or
they may possibly play a role in triggering the deep convection, as discussed below.

Figure 19 shows horizontal plots of high-frequency OLR anomalies during the preconditioning period of MJO1. Convection was enhanced along the equator and propagated eastward over the IO (20°–100°E; Figs. 19a–c). The high-frequency OLR anomalies and high-frequency eddies in the southern part of the IO box may be associated with \((-\mathbf{V}^* \cdot \mathbf{V}^*)\) (Fig. 7c). The passage of high-frequency convective anomalies was concurrent with the intraseasonal convective development at the MJO initialization (Figs. 18e,f and 19b,c). This suggests a triggering effect of the high-frequency disturbances on the convective initiation of MJO1.

According to the time-localized space–time power spectra based on the combined Fourier-wavelet transform (CFWT; Kikuchi 2013), a significant Kelvin wave peak was found in the preconditioning stage of MJO1. Figures 20 and 21 show the horizontal plots and time-longitude section, respectively, of Kelvin wave–filtered (Kiladis et al. 2009) IR Tb. Comparison of Figs. 19 and 20 shows that the eastward-propagating high-frequency signal was primarily interpreted as a convectively coupled Kelvin wave. The amplification of a convectively coupled Kelvin wave around the convective initiation date over the IO is also evident in the time–longitude plot (Fig. 21). Further analysis will be presented elsewhere. The implication here is that convectively coupled high-frequency equatorial waves could act to trigger MJO convective initiation when dry anomalies in the middle troposphere

![Image](https://example.com/fig17.png)

**FIG. 17.** Intraseasonal meridional velocity and wind vector at 600 hPa averaged during 9–15 Oct from ERA-Interim. Square indicates the target IO domain.

![Image](https://example.com/fig18.png)

**FIG. 18.** Horizontal plots of the intraseasonal NOAA OLR and 600-hPa wind (vector) from ERA-Interim at 3-day intervals during 3–18 Oct 2011. Squares indicate the target IO domain.

![Image](https://example.com/fig19.png)
are reduced through horizontal moisture advection. This process will be especially effective in situations where the free atmosphere is dried out and moisture buildup by vertical advection takes a long time, as seen in MJO1.

6. Conclusions

Convective initiation processes in the two MJO events that occurred during the CINDY2011/DYNAMO IOP (MJO1, MJO2) were investigated using ERA-Interim and satellite data. The comparison of the ERA-Interim results with sounding data collected in the field operation and TRMM 3B42v7 precipitation data confirmed reasonable matching, but with dry and cold biases and weaker intraseasonal contrasts of precipitation in ERA-Interim. Moistening processes prior to the convective initiation were examined by budget analysis. The convective initiation date was defined as when the intraseasonal (20–80-day period) OLR anomalies turned negative in the target domain (10°S–10°N, 60°–90°E) over the equatorial Indian Ocean. The initiation dates of MJO1 and MJO2 were 18 October and 18 November 2011, respectively. In both cases, the precipitable water and column-integrated MSE increased during 8–9 days prior to the convective initiation date. These were consistent with previous studies of the climatological MJO (Maloney 2009; Zhao et al. 2013), but with slower buildup of moisture in the CINDY2011/DYNAMO cases, especially for MJO1. The moisture budget analysis that partitioned each variable into low-frequency (<0.80-day period), intraseasonal, and high-frequency (<20-day period) components revealed that 1) zonal advection of low-frequency moisture (maximum over the Maritime Continent) by intraseasonal easterly anomalies and 2) advection of intraseasonal moisture (in the western IO or southern ITCZ) by the basic zonal wind primarily accounted for the moistening in the free atmosphere during the preconditioning period. The drier conditions in the suppressed phase of intraseasonal variability compared with the climatology give one possible
reason for the pronounced midlevel moistening by horizontal advection. In the canonical MJO initiation, low-level advections more smoothly led to moist anomalies and subsequent feedback between latent heat release and upward motion in the deep troposphere (Zhao et al. 2013). Once the MJO-scale convective envelope was established, moistening by vertical advection and drying by meridional advection were pronounced. The latter was essential to the eastward propagation of MJO (Maloney 2009; Kerns and Chen 2014; Kim et al. 2014). These features were also present in the current study.

Interestingly, nonlinear moisture advection associated with high-frequency fluctuations amounted to the same order of magnitude as the primary terms in the preconditioning period of both MJO episodes, especially in the middle tropospheric meridional components. The effects were pronounced over the southern ITCZ around the CINDY2011/DYNAMO sounding array and were likely associated with high-frequency convective disturbances that were abundant during the IOP (JC13; Zuluaga and Houze 2013). Systematic upscale transport of moisture by these disturbances was suggested.

To gain an insight into the effects of high-frequency disturbances on the MJO convective initiation, equatorial waves were examined by the CFWT (Kikuchi 2013) using the 3-hourly 0.5° IR Tb data. In MJO1, high-frequency deep convection developed concurrently with the intraseasonal convective initiation. This is suggestive of the triggering effects of the high-frequency disturbances, which is another research topic regarding the MJO convective initiation. The Kelvin wave–filtered Tb anomalies primarily accounted for the high-frequency signal. L14, using high-resolution multiple reanalysis, also showed a significant contribution of convectively coupled Kelvin waves in the early preconditioning process of MJO1 over the western IO. Because the effects of the high-frequency disturbances depend on the resolvable scale of the analysis data, further investigations using high temporal and spatial resolution products is warranted.

Compared with MJO1, the moistening process in MJO2 occurred more rapidly and spontaneously. The difference was attributable to the more suppressed thermal conditions over the western IO in MJO1 than in MJO2, as well as the seasonal march and activity of equatorial and off-equatorial disturbances. A stronger forcing, such as superposition of multiple disturbances, may have been required to initiate MJO convection in the former than in the latter.

As a possible mechanism of easterly acceleration in the lower to middle troposphere, we propose a contribution of off-equatorial wave activity here. Both MJO cases were accompanied by amplification of an off-equatorial Rossby wave train that was connected to the equatorial zonal wind field. The intrusion of the wave train was more significant in the middle troposphere than in upper and lower troposphere. The circumnavigation of zonal wind anomalies was not as clear in the middle troposphere as in the upper troposphere, suggesting that other processes (e.g., local forcing by extratropical wave disturbances or organized convection) were of greater relevance.

These results imply that extratropical forcing may affect MJO convective initiation through both barotropic energy conversion from upper-tropospheric wave activity, as discussed by previous studies (Zhao et al. 2013; Ray and Li 2013), and also through local moistening in the lower to middle troposphere. Previous case studies demonstrated that the upper-tropospheric Rossby wave train can initiate intraseasonal convective organization over the IO (Hsu et al. 1990), and the cold surge event can help MJO convective initiation through deep convection organization over Madagascar (Wang et al. 2012).

The local processes that lead to MJO convective initiation may be different for different cases. How often and in which situations our findings are applicable is uncertain. Examination of the dependence of local MJO convective initiation processes on the environment (e.g., seasonal and transient distributions of SST, basic zonal wind, stability, and extratropical and equatorial wave activity) will provide perspective.
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