Equatorially Antisymmetric Features in the Initiation Processes of the Madden–Julian Oscillation Observed in Late October during CINDY2011

Qoosaku MOTEKI

Department of Coupled Ocean-Atmosphere-Land Processes Research, Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan

(Manuscript received 4 February 2015, in final form 29 June 2015)

Abstract

In this study, we investigated the equatorially antisymmetric features in the initiation processes of the Madden–Julian oscillation (MJO) event in late October during the cooperative Indian Ocean experiment on intraseasonal variability in the year 2011. This MJO event developed when the thermal equator was drastically shifted from the Northern Hemisphere to the equator from September to October as the Mascarene High over the southern Indian Ocean decayed and shifted southward. The large-scale fields of sea level pressure, temperature, and moisture around the MJO exhibited equatorially antisymmetric features. According to their different features in terms of surface convergence, temperature, and moisture, MJO convection over the Indian Ocean after its onset consisted of four distinct convective components: the southern intertropical convergence zone between 10°S and 0° along the meridional sea surface temperature (SST) gradients (s-ITCZ), the northern-ITCZ at the southern edge of the high SST above 29°C over the Bay of Bengal, the vortex disturbance over the Arabian Sea (VDAS) in association with the zonal SST gradients, and the westward-propagating diurnal convection originating from Sumatra. In particular, the double-ITCZ and VDAS were characterized by a steady, low-level convergence zone along the surface potential temperature gradients forced by the SST. Before the onset of the MJO convection, the double-ITCZ was characterized by cross-equatorial vertical circulation that was baroclinically tilted northward, and s-ITCZ convection was inhibited owing to the strong Mascarene High. After the onset, single, larger-scale upward motion was barotropically formed over the equator because of the equatorward shift of the double-ITCZ. Such changes in the equatorially antisymmetric meridional circulation are relevant for the organization of the MJO convection.

Keywords MJO initiation; CINDY; equatorially antisymmetric features; double-ITCZ; Mascarene High

1. Introduction

The Madden–Julian oscillation (MJO; Madden and Julian 1972, hereafter referred to as MJ72) is a large-scale convection propagating eastward at approximately 5 m s⁻¹ over the equator. Tremendous efforts to conduct observations and numerical modeling pertaining to the MJO have been made (Zhang 2005) because of its significant impacts on global weather and climate (Zhang 2013). However, it is often suggested that environmental conditions in the subtropics and midlatitudes influence the initiation and development of the MJO (Ray and Zhang 2010; Adames et al. 2014; Frederiksen and Lin 2013). Among these recent advances, the mechanisms of the initiation of the MJO remain one of the most important subjects.

Several field campaigns have been conducted...
over the Indian Ocean in recent years to investigate this topic. During the boreal fall in 2006, the field campaign Mirai Indian Ocean cruise for the study of the MJO convection onset (MISMO) was conducted to observe the atmosphere and ocean of the central equatorial Indian Ocean (Yoneyama et al. 2008; Katsumata et al. 2009; Yamada et al. 2010; Yasunaga et al. 2010; Moteki et al. 2011). During the boreal fall and winter in 2011–2012, a coordinated field campaign was conducted consisting of four projects: the cooperative Indian Ocean experiment on intraseasonal variability in the year 2011 (CINDY2011, Yoneyama et al. 2013), dynamics of the MJO (DYNAMO), the atmospheric radiation measurement program MJO investigation experiment, and littoral air-sea process. Hereafter, the campaign is simply referred to as CINDY. The MISMO captured a weak MJO event under the strong Indian Ocean dipole (IOD; Saji et al. 1999). The CINDY campaign captured four notable MJO events under La Niña and neutral IOD.

Johnson and Ciesielski (2013) reported the large-scale environmental properties of the two prominent MJOs that occurred during October and November 2011 from the unique measurements of DYNAMO. The authors noted that the heaviest precipitation over the Indian Ocean was located along the strongest sea surface temperature (SST) gradients. In addition, they described the meridional structure of these MJOs and a shift from a double intertropical convergence zone (ITCZ) in the off-equator before the onset to a broad rainfall maximum near the equator after the onset. In general, a majority of studies on the MJO revealed a structure and propagation in the east–west direction on the basis of the illustration of MJ72.

Meanwhile, the ITCZ transition and the seasonal change in the thermal equator from the boreal summer to winter in the north–south direction may be important factors during the initiation phase of the MJO. In fact, Yoneyama et al. (2013) described meridional changes of the MJO events that exhibited equatorially antisymmetric cloud distributions. Moreover, Tung et al. (2014) emphasized the importance of north–south asymmetry in the MJO convection according to their statistical analysis because the strongest MJO signals in deep convection and precipitation are antisymmetric about the equator. They showed that the antisymmetric case propagated with a slightly slower phase speed than the symmetric case but travels farther east over the western Pacific and suggested that there were fundamental differences in the underlying mechanisms between the symmetric and antisymmetric cases. However, kinematic and thermodynamic fields that form such antisymmetric cloud distributions have not been studied in depth. Thus, the evolution of the dynamical structure forming these antisymmetric cloud distributions should be examined to understand the initiation processes of the MJO.

This study describes the equatorially antisymmetric features of dynamical fields in the initiation processes of the MJO observed in late October during CINDY in detail. This MJO was selected for study because it was the first event in the seasonal change from the boreal summer to winter in 2011. Clear transition processes from a convectively suppressed phase to a mature phase were observed in cloud distributions. In this study, the initiation stages of the MJO event were discriminated depending on the patterns of convection and sea level pressure (SLP) compared with those from the illustration of MJ72 over the Indian Ocean. In addition, to clarify the causes of its antisymmetric features, the MJO convection was divided into four segments based on the surface convergence, surface potential temperature (SPT), and SST. The objective of this study is to reveal the distinct evolution of kinematic and thermodynamic structures for each segment that make antisymmetric cloud distributions in the initiation processes of the MJO.

2. Data

A globally merged infrared radiation brightness temperature (Tb) dataset (archived from September 18, 1999 to the present) was used to depict the evolution of the MJO convection (Janowiak et al. 2001). The data were collected at a spatial resolution of 4 km every 30 min. The daily SSTs used in this study were from the National Oceanic and Atmospheric Administration high-resolution blended analysis with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ from November 1981 to the present (Reynolds et al. 2007). The Japanese 55-year reanalysis for 1958–2012 (JRA55, Ebita et al. 2011) was used to investigate the kinematic and thermodynamic structure of the large-scale environment. The dataset has a $1.25^\circ$ horizontal resolution, 38 levels (the surface and 1–1000 hPa), and 6-h intervals. A majority of the intensive observations from CINDY were transmitted to operational centers to be assimilated in the reanalysis. The assimilation scheme of the four-dimensional variational method was adopted in JRA55 using the forecast model T319/L60 (a spatial resolution of approximately 60 km). Satellite data including ocean surface wind microwave scatterometers and radiances from vertical sounders and micro-
Wave imagers were assimilated such that the quality of JRA55 was improved compared with the existing reanalysis datasets (e.g., Chen et al. 2014). The qualitative patterns of averaged large-scale fields with the JRA55 in this study are largely consistent with other objective analysis datasets (e.g., ERA-Interim, Dee et al. 2011).

3. Large-scale environmental conditions around the MJO convection

Figure 1 shows the time–longitude cross sections of Tb and SLP. The five periods for every 6 days shown in Figs. 1a and 1b are conveniently discriminated in comparison with the illustration of the MJO for the basic features of the developmental stages in Fig. 16 of MJ72 (Fig. 1c). The date of the MJO onset is October 21, 2011 according to previous studies (e.g., Yoneyama et al. 2013). The five periods were named as follows:

P1: convectively suppressed for the period of October 9–14
P2: developing convection just before the MJO onset for the period of October 15–20

Fig. 1. Time–longitude cross sections of (a) Tb (color, K), the surface westerly wind speed (contours for every 1 m s$^{-1}$), and (b) SLP (color, hPa), and the contour for 260 K of Tb averaged for 10°S–10°N from 00Z October 9 to 23Z November 7. The labels “W”, “C”, and “E” represent the western, central, and eastern parts of the MJO over the Indian Ocean, respectively. (c) Illustration of the MJO based on Fig. 16 of MJ72. The shaded zone in (c) indicates the width of the longitudinal sections of (a) and (b).
P3: developing convection just after the MJO onset for the period of October 21–26
P4: mature MJO convection for the period of October 27 to November 1
P5: eastward propagation of the MJO convection for the period of November 2–7

On the basis of the qualitative features of Tb, SLP, and the westerly winds, the MJO convection was divided into the following three zonal parts: western (labeled “W”, 55–70°E), central (labeled “C”, 70–85°E), and eastern (labeled “E”, 85–100°E). Specific features of periods P1–P5 compared with the illustration of the MJO of MJ72 are as follows:

P1 (corresponds with stage E in Fig. 1c): there are nearly no organized convective cloud areas (< 260 K) under high-SLP conditions greater than 1010 hPa over the entire Indian Ocean.

P2 (corresponds with stage F in Fig. 1c): convection with the surface westerly winds in the “W” part appears. In the “E” part, the westward-propagating diurnal convection from Sumatra (WPDS) is gradually intensified and appears through the end of P3. The SLP for all parts of “W”, “C”, and “E” is significantly decreased by approximately 2 hPa compared with P1.

P3 (corresponds with stage G in Fig. 1c): convective activities over the entire Indian Ocean become stronger, and convective cloud areas (< 260 K) appear in all three parts of “W”, “C”, and “E.” The significantly decreased SLP for the same region is less than 1009 hPa.

P4 (corresponds with stage G in Fig. 1c): the MJO convection with the westerly winds in the “C” part is matured and begins to propagate eastward at a speed of approximately 5 m s⁻¹. Prior to the eastward propagation, the decreased SLP region spreads to the east. Although convection during P4 is stronger than that during P3, P4 also corresponds with the same G stage in Fig. 1c as P3 because the location of the convective center and decreased pressure remains over the Indian Ocean.

P5 (corresponds with stage H in Fig. 1c): the MJO convection with the westerly winds is shifted to the land of Indonesia (90–120°E), and the Indian Ocean is again dominated by a high pressure greater than 1010 hPa.

Because this case was the first event in the seasonal change from the boreal summer to winter in 2011, clear transition processes from a convectively suppressed phase (P1) to a mature phase (P5) were observed in cloud distributions. Figure 2 shows the time–latitude cross sections of Tb and SLP averaged for the central part of the MJO labeled as “C” in Fig. 1 and representative horizontal cloud images for P1–P4. During P1 and P2 before the MJO onset, a double ITCZ is observed at 5–15°N and 10°S–0°. With regard to the northern-ITCZ (n-ITCZ) over the land of India to the north of 5°N, the low-SLP area exhibits a diurnal variation with an amplitude of approximately 1 hPa for P1 and P2 (Fig. 2a). The diurnal variation of the n-ITCZ convection could be due to decreased pressure in the evening after land heating in the daytime, as shown by Basu (2007). The southern-ITCZ (s-ITCZ) convection is continuously formed on the northern edge of the Mascarene High (labeled in Figs. 2b-d), which is a subtropical high in the southern Indian Ocean (Rui and Wang 1990; Terray et al. 2003; Fischer et al. 2005; Zhang 2005) and is slightly shifted northward to the equator (Fig. 2a). Zuluaga and Houze (2013) revealed that convective cloud elements observed within the s-ITCZ by a Doppler radar had a periodicity of 2 days, which was associated with a westward-propagating intraseasonal wave. Thus, the meridional pattern of SLP and the periodicity of the convective activities in the double-ITCZ are relatively antisymmetric with respect to the equator.

During P3 just after the MJO onset, the Mascarene High was weakened and shifted to the south (Fig. 2d), and the area with a pressure of less than 1009 hPa is shifted from the land of India to the equatorial Indian Ocean at 10°S–10°N (Fig. 2a). The convective area develops a nearly “equator-symmetric” shape and develops rapidly within the area with a pressure of less than 1009 hPa, as indicated by the bold contour (Fig. 2a). Many studies on the variability of the location and strength of the Mascarene High (Terray et al. 2003; Suzuki et al. 2004; Xue et al. 2004; Fischer et al. 2005; Feng et al. 2013; Manatsa et al. 2014; Morioka et al. 2015) have noted that the Mascarene High influences monsoonal flow and SST variation over the Indian Ocean. According to the large-scale environmental conditions shown here, the weakening and southward shift of the Mascarene High could be one of the favorable environmental conditions enabling the MJO onset to break the convective suppressed period of P1 over the equatorial Indian Ocean. Although few studies have examined the interdependency of the Mascarene High and MJO, Pohl et al. (2009) described the relationship between...
Fig. 2. (a) Time–latitude cross sections of Tb (color, K) and SLP (contours for every 1 hPa) averaged for 70–85°E. The 1009 hPa contour is indicated by the bold black line. The horizontal solid line at 5°N shows the representative location of the southern tip of Sri Lanka. The red rectangle indicates the 1-day mean period for (b)-(e). The 1-day mean fields for Tb (color, K), the SLP (contours for every 2 hPa), and the surface wind vectors are shown on (b) October 12, (c) October 18, (d) October 24, and (e) October 30. The synoptic-scale low- and high-pressure systems are labeled as “L” and “H”, respectively. The dashed rectangles labeled as “n-ITCZ” and “s-ITCZ” roughly show the cloud areas associated with “n-ITCZ” and “s-ITCZ”, respectively, in each panel.
an intraseasonal modulation of the Mascarene High and MJO by applying a principal component analysis. The variations in the environmental conditions and convection during P1 and P2 are consistent with a previous study showing that the Mascarene High generally inhibits convection owing to the trade inversion (Bovalo et al. 2012).

During the mature period of P4, a strong extratropical cyclone appears to the south of 20°S, and its appearance is synchronized with the local minimum pressure of less than 1008 hPa in the equatorial region and the intensification of the MJO convection (Fig. 2c). Ray and Zhang (2010) found that a substantial wave activity propagated northward toward the tropics from the southern Indian Ocean, and the latitudinal transport of the westerly momentum was crucial for the MJO development. The extratropical cyclone that synchronized with the intensification of the MJO observed in the present study could be a reasonable specific candidate for the wave source discussed in their study.

To demonstrate the seasonal variation of the large-scale environmental conditions surrounding the MJO, the time–latitude cross sections from September to December are shown in Fig. 3. The SLP anomaly (SLPA) from the temporal mean is normalized by the standard deviation for the difference in the amplitude of the SLP variation (Fig. 3b) according to the
following calculation:

\[ \text{SLPA} = \frac{\text{SLP} - \text{mean}(\text{SLP})}{\text{std}(\text{SLP})}, \]

\text{mean}(\text{SLP}) and \text{std}(\text{SLP}) are the mean and standard deviation of SLP, respectively, for the period of September 1 to December 31, 2011.

Three MJO events, denoted MJO1, MJO2, and MJO3, were observed during the period from October to December, as reported by Yoneyama et al. (2013) (Fig. 3a). The seasonal migration of the convective centers of the MJOs (indicated by the red dashed arrow) was consistent with that of the high-SST area above 27°C (Figs. 3a, d). The continuous convergence and convection of the s-ITCZ (Fig. 3c) correspond with the northern edge of the northerly winds at the large meridional gradients of the SST and SPT (Fig. 3d). This coincidence of the s-ITCZ convection and meridional gradients is consistent with the features in the averaged field for MJO1 and MJO2, as noted by Johnson and Ciesielski (2013). In contrast, the convergence and SPT gradients of n-ITCZ convection are considerably weaker than those of the s-ITCZ.

All three MJO events are associated with a low pressure of less than 1009 hPa, which corresponds with the negative \( SLPA \) less than \(-1 \sigma \) against the standard deviation for four months (Fig. 3b). During September and the first half of October, the low-pressure area is distributed over the land of India (10°–40°N), and the Mascarene High is dominant over the Indian Ocean to the south of 10°N. The large-scale pair of low- and high-pressure systems, which drive cross-equatorial southerly monsoon flows owing to the meridional pressure gradient, is thought to be caused by the land–sea heating contrast (Turner and Annamalai 2012). Thus, the low-pressure system caused by surface heating is seasonally shifted from the Northern Hemisphere to the Southern Hemisphere as a result of the southward migration of the thermal equator led by the seasonal cycle of solar insolation. The large-scale low-pressure system, which is shifted southward across the equatorial Indian Ocean, could represent favorable environmental conditions for the development of the three MJO events.

In addition, the low \( SLPA \) associated with the MJOs in Fig. 3b extends to the midlatitudes in the Southern Hemisphere. The low \( SLPA \) in the southern Indian Ocean is associated with the extratropical cyclone that passes eastward, which is same situation depicted in Fig. 2e. Thus, MJO1, MJO2, and MJO3 are associated with the passages of extratropical cyclones over the southern Indian Ocean. This feature further suggests that the influence of extratropical cyclones is important for the MJO development, as argued by recent studies (e.g., Ray and Zhang 2010; Adames et al. 2014; Kerns and Chen 2014). In addition, this feature is consistent with the results of Zhang and Dong (2004), who demonstrated the latitudinal seasonal migration of the MJO signal across the equator.

4. Kinematic and thermodynamic structures of the MJO at each stage

Section 3 shows that the evolution of large-scale dynamical fields surrounding the MJO is different between the Northern and Southern Hemispheres, and variation of the Mascarene High is one of the important environmental factors triggering the MJO. This section describes the kinematic and thermodynamic structures of the MJO itself at P1 (convectively suppressed), P2 (just before the MJO onset), P3 (just after the MJO onset), and P4 (the mature MJO) as the initiation processes. As shown in Fig. 3, the seasonal migrations of SLP and surface convergence largely corresponded with those of the SST and the meridional gradient of the SPT. Here, their horizontal distributions, averaged for each development stage of P1–P4, are described.

Figure 4 shows the horizontal distributions of \( T_b \), SLP, surface convergence, and surface winds. The four rectangles of the northwestern (NW), southwestern (SW), central (C), and eastern (E) segments of the MJO convection are determined depending on convection (\( T_b < 280 \) K) and the local maximum of the surface convergence (> \( 1 \times 10^{-6} \) s\(^{-1}\)) at the stage just after the onset in Fig. 4c. NW and SW are associated with a pair of cyclonic vortices, C is the center of the most active convection with the convergence ahead of the westerly winds, and E is WPDS (Fig. 1a) in association with the convergence of the easterly winds. During P1 (Fig. 4a), the weak s-ITCZ convection associated with the surface convergence zone between 10°S and 0° is located at the northern edge of the southeasterly monsoon flow from the strong Mascarene High. The n-ITCZ convection is extended from the coast of India to the southern Bay of Bengal. During P2 (Fig. 4b), the SLP over the equatorial Indian Ocean decreases slightly in association with the weakening and southward shift of the Mascarene High. The s-ITCZ convection develops rapidly and enlarges horizontally within the decreased SLP over the equator. In the Northern Hemisphere, an isolated cyclonic vortex disturbance over the Arabian
Fig. 4. Horizontal distributions of Tb (color, K), SLP (contours for every 0.5 hPa), the surface convergence (red dashed contours for 1, 2, 4, and \(8 \times 10^{-6} \text{s}^{-1}\)), and the surface winds (vector) averaged for (a) P1, (b) P2, (c) P3, and (d) P4. The synoptic scale low- and high-pressure systems are labeled as “L” and “H”, respectively. The intensive upper air sounding sites and research vessels involved in the CINDY campaign are shown by the white and red circles, respectively. The four dotted rectangles, labeled as “NW”, “SW”, “C”, and “E” in (c), show the areas where averages were calculated for Fig. 8.
Sea (VDAS) with a scale of approximately 1000 km appears to the west of the n-ITCZ. During P3 (Fig. 4c), the center of the low-pressure system is over the equator, and the large-scale convection of the MJO develops: convective areas, which have zonal and meridional widths of approximately 7000 and 2000 km, respectively, are formed with an “equator-symmetric” feature over the Indian Ocean. The pattern of the surface winds within the low-pressure area resembles the Matsuno–Gill pattern of the heat source response in the tropics (Matsuno 1966; Gill 1980). During P4 (Fig. 4d), the low-pressure system of the MJO is separated into two parts over the Arabian Sea and Maritime Continent, and the surface westerly winds over the equator are intensified. In the Northern Hemisphere, a remarkable high-pressure system appears over the subtropics from the land of India to the Indochina Peninsula. Due to such SLP patterns, the flows from both the Northern and Southern Hemispheres toward the equatorial area are considerable.

Through the four periods of P1–P4, the s-ITCZ is found to be maintained at the northern edge of the southeasterly monsoon flow at approximately 10°S–0° in segments SW, C, and E. In other words, the intensification and horizontal enlargement of the s-ITCZ convection are among the essential initiation processes required for the onset of the MJO event. However, in the Northern Hemisphere, the horizontal wind patterns and SLP significantly change at each stage of P1–P4. During the periods before the onset (Figs. 4a, b), the flow from the Arabian Sea to the Bay of Bengal along the southern edge of the low-pressure system is dominant. During the periods after the onset (Figs. 4c, d), the northeasterly outflow from the high-pressure system in the subtropics toward the equatorial area becomes dominant in the Northern Hemisphere. Such drastic changes in wind direction in the Northern Hemisphere could be among the important causes of the intensification of the large-scale convergence over the equator, inducing the organization of the large-scale convective area of the MJO event.

Figure 5 shows the horizontal distributions of the SST, surface atmospheric temperature (SAT), and surface winds. Through P1–P4, the distribution of the SAT is largely consistent with that of the SST; however, its values are 1–2°C lower than those of the SST. Thus, the lower atmosphere is thermally forced by the SST. Consequently, the significant meridional gradients of the SST (1–4°C/1000 km) coincide with those of the SPT, as shown in Fig. 6. In particular, the local maximum of the meridional gradients of the SPT along the s-ITCZ coincides well with that of the SST through P1–P4. Hereafter, the surface convergence zone associated with convection along the meridional gradients of the SST in the 6-day average field is conveniently defined as a “stationary front.”

During the periods of P1-P2 before the MJO onset, separate convective areas are formed along the stationary fronts rather than the maximum SST (Figs. 4a, b, 5a, b). In particular, the stationary front of the s-ITCZ between 10°S and 0° corresponds well with the remarkable meridional gradients of the SST. The location of the SST gradients also corresponds with the Seychelles–Chagos thermocline ridge region (Hermes and Reason 2008; Vialard et al. 2009). In addition, the relationship between the high (low) SLP in Fig. 4 and the low (high) SST in Fig. 5 supports the notion that the surface convergence of the s-ITCZ is due to the pressure adjustment mechanism (Lindzen and Nigam 1987; Kessler and Kleeman 2000; Pezzi et al. 2009). The pressure adjustment mechanism is a well-known explanation for the cause of the surface convergence over the tropics: an SST-induced SAT difference hydrostatically adjusts the pressure gradient that drives the surface winds to produce the surface convergence.

The stationary front of the n-ITCZ between 5° and 15°N during P1 and P2 also corresponds well with the remarkable meridional gradients of the SST at the southern edge of the warmest SST area, which is above 29°C over the Bay of Bengal, in association with the pressure adjustment mechanism (Figs. 5a, b). During P3 and P4, the stationary front of the n-ITCZ is separated from the SST gradients and shifted southward (Figs. 6c, d). Moreover, the n-ITCZ is formed between warmer and cooler SAT areas to the north and south (Figs. 5c, d). That is, the behaviors and SPT gradients associated with the n- and s-ITCZ are equatorially antisymmetric. In addition, VDAS in the NW segment is associated with the zonal gradients of SST/SAT at the western edge of the low-pressure system, and the equatorially antisymmetric pattern of the temperature fields in the NW and SW segments is relatively significant.

During the MJO onset of P3, the stationary fronts of both the n- and s-ITCZ are shifted toward the equator. In such a situation, the separated convective areas during P1 and P2 (Figs. 4a, b) could be easily concentrated within the decreased SLP over the equator, and a well-organized convection on a larger scale could easily develop (Fig. 4c). In addition, the westerly winds toward the center of the low-pressure
Fig. 5. Horizontal distributions of the SST (color, °C), SAT (contours for every 0.2°C > 27°C, bold contours for every 1°C), and the surface winds (vector) averaged for (a) P1, (b) P2, (c) P3, and (d) P4. The convergence zone with the large contrast in the SPT is indicated by the symbol for the stationary front associated with convection, which is referred to in Figs. 4–6.
Fig. 6. Horizontal distribution of the meridional gradient of the SST (color, K/1000 km), the SPT (contours for 1, 2, and 4 K/1000 km), and the surface winds (vector) averaged for (a) P1, (b) P2, (c) P3, and (d) P4.
system become more prominent over the high SST of above 29°C (Fig. 5c). The surface zonal convergence ahead of the westerly winds is intensified in the C segment and is thought to contribute to the organization of the convection (Fig. 4c).

Moreover, with regard to the initiation of convection, it is necessary to investigate the relationship between the dynamical and moisture fields below the level of condensation (approximately 500 m: 950 hPa). Figure 7 shows the horizontal distributions of the specific humidity, vertical vorticity, and winds at the surface. Through P1–P4, the local surface specific humidity maxima along the s-ITCZ correspond well with the local minimums of the negative vertical vorticity. In other words, the specific humidity contrast does not necessarily correspond with the symbols of the stationary front that is defined by the SPT gradient and convergence, and the specific humidity tends to be high within the cyclonic vortex. In particular, the local minimums of the vorticity in the SW and E segments are quasi-stationary through P1–P4, and they could contribute to the formation of the s-ITCZ convection and WPDS.

In contrast, the cyclonic vortex of VDAS in the NW segment moves southward toward the equator during P1–P3 (Figs. 7a, b, c) and turns northwestward during P4 (Fig. 7d). The distribution of the specific humidity over the Arabian Sea changes considerably in association with the motion of VDAS. Thus, cyclonic-vortices in the NW and SW segments have different motions and moisture variations, although they apparently form an equator-symmetric pair during P2–P4. With regard to the n- and s-ITCZ, the surface specific humidity over the Bay of Bengal is the highest in the analyzed region, and the surface moisture field exhibits considerable antisymmetry with respect to the equator.

To summarize Figs. 4–7, the MJO convection is divided into four convective components based on the differences in thermal environmental features and initiation triggers. That is, WPDS over the eastern Indian Ocean, the s-ITCZ convection between 10°S and 0°, the n-ITCZ convection over the Bay of Bengal, and the VDAS convection clearly exhibit different behaviors. The MJO onset during P3 is induced by a process in which the four convective components become unified within a large-scale low-pressure system over the equatorial Indian Ocean.

To summarize the features in each segment at each stage shown in Figs. 4–7, Fig. 8 shows the time series of the Tb anomaly, surface divergence, surface vertical vorticity, SLP4, and surface specific humidity averaged in the NW, SW, C, and E segments. Here, the anomalies from the temporal average for P1–P5 are used to illustrate differences in the features at each stage.

Features of the negative SLP4 of 1–2 hPa and the positive specific humidity anomaly of 0.5–1 g kg\(^{-1}\) are common for all segments during P3 and P4. In addition, a continuous decreasing trend in the SLP through P1–P3 is shown in all segments, and the SLP decreases prior to the development of convection indicated by Tb. This pattern suggests that the large-scale SLP decreases in P1–P3 are mainly due to the weakening and southward shift of the Mascarene High rather than to diabatic heating by convection. The large-scale mean updraft could become large in association with continuous SLP decreases exceeding 10 days and with the moistening of the lower troposphere by shallow convection over a time scale of a few days, which was discussed in preconditioning studies (Waite and Khouider 2010; Masunaga 2013; Bellenger et al. 2015), and could be easily and gradually activated under conditions involving continuous SLP decreases.

In contrast, the temporal variations in the Tb anomaly, divergence, and vorticity are significantly different among the segments. Although the local minima of the negative Tb anomalies less than −20 K in the NW, C, and E segments appear in periods P3, P4, and P5 (Figs. 8a, c, d), respectively, the variation in the Tb in the SW segment (Fig. 8b) is clearly smaller than in the other three segments because convection in the SW segment is continuously maintained by the steady s-ITCZ convergence through P1–P5. In the C segment (Fig. 8c), the representative MJO convection for P3 and P4 is characterized by the variation in Tb and the shift of the double-ITCZ toward the equator (Figs. 3c, 4–7). In the E segment (Fig. 8d), the standard deviation of Tb (25.9 K) is clearly larger than in the other three segments (NW: 23.3 K, SW: 18.7 K, C: 24.7 K). This larger standard deviation could be due to the influence of the diurnal variation in WPDS, as shown in Fig. 1a, although diurnal variation cannot be directly identified in the box average of Fig. 8d. In the NW segment (Fig. 8a), the cyclonic vorticity of VDAS is monotonically intensified in association with the continuous decrease in the SLP, and the intensification of the convergence corresponds well with the increase in moisture. According to these features in each segment, although the variations in the C and E segments are consistent with the development and eastward propagation of the MJO, the variations associated with the s-ITCZ and VDAS in the SW and NW segments are relatively
Fig. 7. Horizontal distribution of the specific humidity (color, g kg$^{-1}$), the vertical vorticity (contours for every 2 × 10$^{-6}$ s$^{-1}$), and the surface winds (vector) averaged for (a) P1, (b) P2, (c) P3, and (d) P4.
unaffected by those of the MJO.

Figure 9 shows the meridional structure averaged for the C and E segments, which indicates the representative variations of the MJO convection and decrease in SLP. The positive anomaly of the specific humidity from the meridional average at each pressure level is used as an index of the depth of convective activity. Smaller Tb (plotted at the top panel) values are consistent with deeper moist layers. The difference in vertical circulation between the periods before (P1-P2) and after (P3-P4) the onset is remarkable. For P1 and P2 before the MJO onset (Figs. 9a, b), the two separated cores of the positive moisture anomalies, which are described as a double-ITCZ pattern in Johnson and Ciesielski (2013), are present in the Southern and Northern Hemispheres. However, the vertical structure of the convergence/divergence clearly exhibits antisymmetry with respect to the equator before the onset, and the center of the northward-tilted vertical circulation is located in the Northern Hemisphere. For P3 and P4 after the MJO onset (Figs. 9c, d), the equatorially symmetric core of the positive moisture anomaly and vertical circulation is formed. Thus, the equatorially symmetric strong upward motion of the MJO is formed as a result of the shift of both s- and n-ITCZ toward the equator, as indicated by the triangles at the bottom of each panel.

One of the key features of the initiation processes of
the MJO is this change from the baroclinic structure of the northward-tilted vertical circulation to the barotropic structure. Consequently, the s-ITCZ convection during P4 is greatly intensified and enlarged, as shown by the local minimum Tb at approximately 4°S (Fig. 9d) and a horizontal distribution of Tb of less than 260 K (Fig. 4d). In fact, the low-level convergence and upper-level divergence associated with the s-ITCZ are considerably stronger than those associated with the n-ITCZ during P4 (Fig. 9d). Although the cloud areas of the double-ITCZ merge, the local maxima and minima of divergence associated with the double-ITCZ remain after the MJO onset just above the symbols of the stationary front, as indicated by the triangles at the bottom of each panel.

Figure 10 shows the zonal structure averaged between 7.5°S and 7.5°N. The zonal antisymmetric structure of a typical MJO, as shown in previous studies (Zhang 2005; Kiladis et al. 2009), is exhibited at the mature stage of P4 (Fig. 10d). That is, the low-level convergence below 500 hPa demonstrates a westward tilt with respect to the convection center corresponding with the C and E segments, and convection indicated by the positive moisture anomaly is shallower to the east than to the west. Such a typical zonal structure of the MJO remains unclear at the stage just after the onset at P3, and there are several separated cores of the upper-level divergence in the W, C, and E segments (Fig. 10c). Going backward in time, the origin of convection in the E segment is the diurnal convection over the island of Sumatra (Figs. 1a, 10a, b). This WPDS has been studied by Mori et al. (2004) and Fujita et al. (2011). The authors noted that the diurnal convection over Sumatra migrated up to 400 km from the coastline under the low-level westerly winds and upper-
level easterly winds and was intensified during the active phase of the MJO. The WPDS in the E segment before the MJO onset could be organized by the migration mechanism presented by Mori et al. (2004) and could be intensified as a result of the combination of the southward-shifting n-ITCZ and the expanding s-ITCZ during P3 and P4. As a result of the expansion of the horizontal scale for the low-level convergence/upper-level divergence due to the combination of the convective components, the well-organized zonal circulation of the MJO is believed to be formed during P4.

5. Conclusions

In this study, we documented the initiation processes of an MJO event in late October during CINDY, focusing on its equatorially antisymmetric features. The five developmental stages of the event are conveniently discriminated according to the qualitative features of Tb and SLP compared with the illustration of MJ72. The five periods, denoted P1 (convectively suppressed for the period of October 9–14), P2 (developing convection just before the MJO onset for the period of October 15–20), P3 (developing convection just after the MJO onset for the period of October 21–26), P4 (the mature MJO convection for the period of October 27 to November 1), and P5 (the beginning of the eastward propagation of the MJO convection for the period of November 2–7), corresponded well with each developmental stage defined in the original MJ72 illustration. In this study, we addressed the evolution of the MJO from P1 to P4 of the initiation processes.

On the basis of differential features in terms of surface convergence, temperature, and moisture, MJO convection over the Indian Ocean during its initiation
consisted of four distinct convective components, the s-ITCZ, n-ITCZ, VDAS, and WPDS. In particular, the double-ITCZ and VDAS were clearly characterized by the steady low-level convergence zone along SPT gradients forced by the SST. This MJO event was initiated by a combination of the four convective components, which were distinguished by distinct dynamical generating conditions. This concept of the four convective components in the initial MJO is the most important notion presented in this case study.

Figure 11 shows illustrations of the surface horizontal pattern and the meridional and zonal vertical structures of the MJO event.

P1 (the period convectively suppressed during October 9–14 corresponding with stage E in MJ72): Convection of the s-ITCZ, n-ITCZ, and WPDS was suppressed under conditions in which a strong Mascarene High was dominant over the entire Indian Ocean (Fig. 11a). Although the double-ITCZ convergence was associated with the meridional SPT gradients forced by the SST gradients, which satisfied the relationship with the pressure adjustment mechanism, it exhibited equatorially antisymmetric features in terms of its orientations, periodicities of convection, and motions. The n-ITCZ convection showed that the diurnal cycle was affected by the land of India (see Fig. 2), and the s-ITCZ convection had a longer periodicity of more than 2 days according to Zuluaga and Houze (2013). The warmest and most moist area was distributed over the Bay of Bengal to the north of the n-ITCZ but not over the equator (see Fig. 5), and the baroclinic structure of the cross-equatorially northward-tilted vertical circulation was considerable (Fig. 11c).

P2 (the period of developing convection just before the MJO onset during October 15–20 corresponding with stage F in MJ72): The s-ITCZ convection developed owing to the decreased pressure over the equatorial region in association with the decay of the southward-shifting Mascarene High (Figs. 11b, g). The VDAS convection formed at the western edge of the low-pressure system, and the WPDS gradually intensified in the decreased SLP.

P3 (the period of developing convection just after the MJO onset during October 21–26 corresponding with stage G in MJ72): Convection of the n- and s-ITCZ substantially shifted toward the equator (Fig. 11c), and a large-scale, strong upward motion over the equator was formed (Fig. 11i). Convection of VDAS and WPDS was considerably intensified compared with that during P2 within the large-scale decrease in pressure over the equatorial Indian Ocean. MJO convection was initiated as a result of the combination of the intensified convective components of the VDAS, WPDS, and double-ITCZ. The zonal vertical circulation of the typical MJO shown by previous studies remained poorly organized (Fig. 11j). Although horizontal low-level flows around the large-scale low-pressure system, which resembled the Matsuno–Gill pattern (Fig. 4c), and the meridional vertical circulation of the MJO (Fig. 11i) exhibited apparently equator-symmetric features, the equatorially antisymmetric features of the temperature and moisture fields were important at the surface because the behaviors of the n- and s-ITCZ were completely different from each other.

P4 (the period of mature MJO convection from October 27 to November 1, corresponding with stage G in MJ72): The MJO convection matured in association with the intensified equatorial westerly winds toward the decreased pressure that expanded to the east over the Maritime Continent (Fig. 11d). The mature MJO convection was synchronized with the passage of the extratropical cyclone in the Southern Hemisphere (see Figs. 2, 3). The MJO convection over the eastern Indian Ocean developed considerably owing to the combination of intensified convergence associated with WPDS and double-ITCZ (see Fig. 8d). The VDAS convection moved northwestward, and the synoptic-scale low-pressure system of VDAS was separated from the larger-scale low-pressure system of the MJO. In contrast, the vortex within the western part of the s-ITCZ was stagnant in the areas of 10°S–0° and 50–70°E. That is, the twin-vortex between 50–70°E demonstrated equatorially antisymmetric features even though it exhibited equatorially symmetric features in horizontal circulation and convective activities. The meridional vertical circulation of the double-ITCZ was continuously intensified (Fig. 11k), and the zonal vertical circulation, which resembled that of the typical MJO shown by previous studies, was consequently well organized (Fig. 11 l).
Fig. 11. Illustrations of the initiation processes of the MJO event in late October during CINDY. Horizontal views for (a) P1, (b) P2, (c) P3, and (d) P4; meridional vertical views for (e) P1, (g) P2, (i) P3, and (k) P4; and zonal vertical views for (f) P1, (h) P2, (j) P3, and (l) P4 are shown. In (a), (b), (c), and (d), the synoptic-scale low- and high-pressure systems are labeled by the symbols “L” and “H”, surrounded by bold black lines. The red and blue arrows represent the warmer and colder flows at the surface, respectively. The symbols for the stationary front show the convergence line with remarkable contrast to SPT, which is associated with convection. The cloud areas (Tb < 280 K) are surrounded by the gray dashed lines. The cloud areas of the northern ITCZ, southern ITCZ, vortex disturbance over the Arabian Sea, and westward-propagating diurnal convection originating from Sumatra are labeled as “n-ITCZ”, “s-ITCZ”, “VDAS”, and “WPDS”, respectively. The area of SST >28°C over the Indian Ocean is shaded in pink. In the right panels, convective clouds and the surrounding vertical circulations are represented by gray shaded areas and streamlines, respectively. In (e), (g), (i), and (k), the triangles at the bottom of each panel labeled as “n-ITCZ” and “s-ITCZ” show the locations of the maximum gradients of SPT associated with the n- and s-ITCZ, respectively. In (f), (h), (j), and (l), the labels at the bottom show the convective components of “n-ITCZ”, “s-ITCZ”, “VDAS”, and “WPDS”, respectively, considered in each panel.
The initiation processes from P1 to P4 show that a change in the equatorially antisymmetric meridional circulation affects the organization of the MJO convection. Before the onset, the double-ITCZ exhibited a cross-equatorial vertical circulation that was baroclinically tilted northward, and the s-ITCZ convection was inhibited owing to the strong Mascarene High. According to the large-scale evolution of environmental conditions, the weakening and southward shift of the Mascarene High could be one of the significant key factors triggering this MJO. After the onset, a single larger-scale upward motion was barotropically formed over the equator because of the equatorward shift of the double-ITCZ circulation in association with the weakening of the Mascarene High.

In the initiation processes of this October MJO that consisted of four distinct convective components, the evolution of convective activities in the Southern and Northern Hemispheres was significantly different and demonstrated equatorially antisymmetric features. Previous studies (Kerns and Chen 2014; Nasuno et al. 2015) noted that the October MJO had several different features compared with the MJO observed in November 2011. Kerns and Chen (2014) revealed that dry air intrusions from the Southern Hemisphere to the equatorial region played a key role in the evolution of the November MJO. Nasuno et al. (2015) remarked that the moistening process in the November MJO occurred more rapidly because of more suppressed thermal conditions over the western Indian Ocean in the October MJO than in the November MJO.

The differences between the October and November MJOs shown by previous studies were consistent with the different thermodynamic conditions at the surface in association with the seasonal migration of the thermal equator, as shown in Fig. 3. The October MJO was initiated when the large-scale low-pressure system caused by surface heating was drastically shifted from the Northern Hemisphere to the equator as a result of the decay of the Mascarene High. In contrast, the low-pressure system of the November MJO exhibited a nearly equatorially symmetric distribution (see Fig. 3a). According to the illustration by Kerns and Chen (2014), the moisture field around the November MJO was more equator-symmetric than that around the October MJO shown in Fig. 7.

One of the important features shared by both MJOs was the fact that the passage of the extratropical cyclone in the Southern Hemisphere was synchronized with the MJO development (see Fig. 3b). This synchronicity of the passage of the extratropical cyclone was also confirmed for another MJO observed in December 2011. This finding supports the notion that environmental conditions in the subtropics and midlatitudes influence the initiation and development of the MJOs (Ray and Zhang 2010; Adames et al. 2014; Frederiksen and Lin 2013; Fukutomi and Yasunari 2014). Furthermore, the features of the convective components of the double-ITCZ and WPDS were shared by the three MJOs in 2011.

The common and contrasting features of the individual MJO events discussed here are expected to be considered in statistical analyses using filtering and composite methods (e.g., Wheeler and Weickmann 2001; Wheeler and Hendon 2004; Kiladis et al. 2009). More studies of equatorially symmetric and antisymmetric types of MJOs, as emphasized by Tung et al. (2014), are necessary to understand the initiation processes in consideration of the convective components. In addition, the influence of the passage of extratropical cyclones over the southern Indian Ocean, which synchronized with the October, November, and December MJOs in 2011, should be specifically investigated in the future.

Acknowledgments

The authors acknowledge Drs. Kunio Yoneyama, Takuya Hasegawa, and Takanori Horii of JAMSTEC and Atsuyoshi Manda of Nagasaki University for their valuable suggestions. Thanks are also given to the reviewers and the editor, Dr. Takemi of Kyoto University, for their helpful recommendations. This work is supported in part by MEXT through a Grant-in-Aid for Scientific Research in Innovative Area 2205 and the Japan Society for the Promotion of Science Grants-in-Aid for Scientific Research 22244057.

References

Adames, Á. F., J. Patoux, and R. C. Foster, 2014: The contribution of extratropical waves to the MJO wind field. J. Atmos. Sci., 71, 155–176.
Basu, B. K., 2007: Diurnal variation in precipitation over India during the summer monsoon season: Observed and model predicted. Mon. Wea. Rev., 135, 2155–2167.
Bellenger, H., K. Yoneyama, M. Katsumata, T. Nishizawa, K. Yasunaga, and R. Shirooka, 2015: Observation of moisture tendencies related to shallow convection. J. Atmos. Sci., 72, 641–659.
Bovalo, C., C. Barthe, and N. Bègue, 2012: A lightning climatology of the South-West Indian Ocean. Nat. Hazards Earth Syst. Sci., 12, 2659–2670.
Chen, G. X., T. Iwasaki, H. L. Qin, and W. M. Sha, 2014: Evaluation of the warm-season diurnal variability over East Asia in recent reanalyses JRA-55, ERA-Interim, NCEP CFSR, and NASA MERRA. J. Climate, 27, 5517–5537.

Dee, D. P., S. M. Uppala, A. J. Simmons, P. Berrisford, P. Poli, S. Kobayashi, U. Andrae, M. A. Balsomeda, G. Balsamo, P. Bauer, P. Bechtold, A. C. M. Beljaars, L. van de Berg, J. Bidlot, N. Bormann, C. Delsol, R. Dragani, M. Fuentes, A. J. Geer, L. Haimberger, S. B. Healy, H. Hersbach, E. V. Holm, L. Isaksen, P. Kallenber, M. Kohler, M. Matricardi, A. P. McNally, B. M. Monge-Sanz, J. J. Morcrette, B. K. Park, C. Peubey, P. de Rosnay, C. Tavolato, J. N. Thepaut, and F. Vitart, 2011: The ERA-Interim reanalysis: configuration and performance of the data assimilation system. Quart. J. Roy. Meteor. Soc., 137, 553–597.

Ebita, A., S. Kobayashi, Y. Ota, M. Moriya, R. Kumabe, K. Onogi, Y. Harada, S. Yasui, K. Miyaoka, K. Takahashi, H. Kamahori, C. Kobayashi, H. Endo, M. Soma, Y. Okawa, and T. Ishimizu, 2011: The Japanese 55-year reanalysis “JRA-55”: An interim report. SOLA, 7, 149–152.

Feng, J., J. Li, and H. Xu, 2013: Increased summer rainfall in northwest Australia linked to southern Indian Ocean climate variability. J. Geophys. Res. Atmos., 118, 467–480.

Fischer, A. S., P. Terray, E. Guilyardi, S. Gualdi, and P. Delecluse, 2005: Two independent triggers for the Indian Ocean dipole/zonal mode in a coupled GCM. J. Climate, 18, 3428–3449.

Frederiksen, J. S., and H. Lin, 2013: Tropical–extratropical interactions of intraseasonal oscillations. J. Atmos. Sci., 70, 3180–3197.

Fujita, M., K. Yoneyama, S. Mori, T. Nasuno, and M. Satoh, 2011: Diurnal convection peaks over the eastern Indian Ocean off Sumatra during different MJO phases. J. Meteor. Soc. Japan, 89A, 317–330.

Fukutomi, Y., and T. Yasunari, 2014: Extratropical forcing of tropical wave disturbances along the Indian Ocean ITCZ. J. Geophys. Res. Atmos., 119, 1154–1171.

Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. Quart. J. R. Meteor. Soc., 106, 447–462.

Hermes, J. C., and C. J. C. Reason, 2008: Annual cycle of the South Indian Ocean (Seychelles-Chagos) thermocline ridge in a regional ocean model. J. Geophys. Res. Oceans, 113, C04035, doi:10.1029/2007JC004363.

Janowiak, J. E., R. J. Joyce, and Y. Yarosh, 2001: A real-time global half-hourly pixel-resolution infrared dataset and its applications. Bull. Amer. Meteor. Soc., 82, 205–217.

Johnson, R. H., and P. E. Ciesielski, 2013: Structure and properties of Madden-Julian oscillations deduced from DYNAMO sounding arrays. J. Atmos. Sci., 70, 3157–3179.

Katsumata, M., R. H. Johnson, and P. E. Ciesielski, 2009: Observed synoptic-scale variability during the developing phase of an ISO over the Indian Ocean during MISMO. J. Atmos. Sci., 66, 3434–3448.

Kerns, B. W., and S. S. Chen, 2014: Equatorial dry air intrusion and related synoptic variability in MJO initiation during DYNAMO. Mon. Wea. Rev., 142, 1326–1343.

Kessler, W. S., and R. Kleemann, 2000: Rectification of the Madden-Julian oscillation into the ENSO cycle. J. Climate, 13, 3560–3575.

Kiladis, G. N., M. C. Wheeler, P. T. Haertel, K. H. Straub, and P. E. Roundy, 2009: Convectively coupled equatorial waves. Rev. Geophys., 47, RG2003, doi:10.1029/2008RG000266.

Lindzen, R. S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the tropics. J. Atmos. Sci., 44, 2418–2436.

Madden, R. A., and P. R. Julian, 1972: Description of global-scale circulation cells in tropics with a 40–50 day period. J. Atmos. Sci., 29, 1109–1123.

Manatsa, D., Y. Morioaka, S. Behera, C. Matarita, and T. Yamagata, 2014: Impact of Mascarene High variability on the East African ‘short rains’. Climate Dyn., 42, 1259–1274.

Masunaga, H., 2013: A satellite study of tropical moist convection and environmental variability: A moisture and thermal budget analysis. J. Atmos. Sci., 70, 2443–2466.

Matsuno, T., 1966: Quasi-geostrophic motions in the equatorial area. J. Meteor. Soc. Japan, 44, 25–43.

Mori, S., J.-I. Hamada, Y. I. Taudid, M. D. Yamanaka, N. Okamoto, F. Murata, N. Sakurai, H. Hashiguchi, and T. Sribimawati, 2004: Diurnal land-sea rainfall peak migration over Sumatera Island, Indonesian maritime continent, observed by TRMM satellite and intensive rawinsonde soundings. Mon. Wea. Rev., 132, 2021–2039.

Moroioka, Y., K. Takaya, S. K. Behera, and Y. Masumoto, 2015: Local SST impacts on the summertime Mascarene high variability. J. Climate, 28, 678–694.

Moteki, Q., K. Yoneyama, R. Shirooka, H. Kubota, K. Yasunaga, J. Suzuki, A. Seiki, N. Sato, T. Enomoto, T. Miyoshi, and S. Yamane, 2011: The influence of observations propagated by convectively coupled equatorial waves. Quart. J. Roy. Meteor. Soc., 137, 641–655.

Nasuno, T., T. Li, and K. Kikuchi, 2015: Moistening processes before the convective initiation of Madden–Julian oscillation events during the CINDY2011/ DYNAMO period. Mon. Wea. Rev., 143, 622–643.

Pezzi, L. P., R. B. de Souza, O. Acevedo, I. Wainer, M. M. Mata, C. A. E. Garcia, and R. de Camargo, 2009: Multiyear measurements of the oceanic and atmospheric boundary layers at the Brazil-Malvinas
confluence region. *J. Geophys. Res. Atmos.*, 114, D19103, doi:10.1029/2008JD011379.

Pohl, B., N. Fauchereau, Y. Richard, M. Rouault, and C. J. C. Reason, 2009: Interactions between synoptic, intraseasonal and interannual convective variability over Southern Africa. *Climate Dyn.*, 33, 1033–1050.

Ray, P., and C. Zhang, 2010: A case study of the mechanics of extratropical influence on the initiation of the Madden–Julian oscillation. *J. Atmos. Sci.*, 67, 515–528.

Reynolds, R. W., T. M. Smith, C. Liu, D. B. Chelton, K. S. Casey, and M. G. Schlax, 2007: Daily high-resolution-blended analyses for sea surface temperature. *J. Climate*, 20, 5473–5496.

Rui, H., and B. Wang, 1990: Development characteristics and dynamic structure of tropical intraseasonal convection anomalies. *J. Atmos. Sci.*, 47, 357–379.

Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian Ocean. *Nature*, 401, 360–363.

Suzuki, R., S. K. Behera, S. Iizuka, and T. Yamagata, 2004: Indian Ocean subtropical dipole simulated using a coupled general circulation model. *J. Geophys. Res. Oceans*, 109, C09001, doi:10.1029/2003JC001974.

Terray, P., D. Delecluse, S. Labattu, and L. Terray, 2003: Sea surface temperature associations with the late Indian summer monsoon. *Climate Dyn.*, 21, 593–618.

Tung, W.-W., D. Giannakis, and A. J. Majda, 2014: Symmetric and antisymmetric convection signals in the Madden–Julian oscillation. Part I: Basic modes in infrared brightness temperature. *J. Atmos. Sci.*, 71, 3302–3326.

Turner, A. G., and H. Annamalai, 2012: Climate change and the South Asian summer monsoon. *Nat. Climate Change*, 2, 587–595.

Vialard, J., J. P. Duvel, M. J. McPhaden, P. Bourouet-Aubertot, B. Ward, E. Key, D. Bourras, R. Weller, P. Minnett, A. Weill, C. Cassou, L. Eymard, T. Fristedt, C. Basdevant, Y. Dandonneau, O. Duteil, T. Izumo, C. de Boyer Montegut, S. Masson, F. Marsac, C. Menkes, and S. Kennan, 2009: CIRENE air-sea interactions in the Seychelles-Chagos thermocline ridge region. *Bull. Amer. Meteor. Soc.*, 90, 45–61.

Waite, M. L., and B. Khouider, 2010: The deepening of tropical convection by congestus preconditioning. *J. Atmos. Sci.*, 67, 2601–2615.

Wheeler, M., and K. M. Weickmann, 2001: Real-time monitoring and prediction of modes of coherent synoptic to intraseasonal tropical variability. *Mon. Wea. Rev.*, 129, 2677–2694.

Wheeler, M. C., and H. H. Hendon, 2004: An all-season real-time multivariate MJO index: Development of an index for monitoring and prediction. *Mon. Wea. Rev.*, 132, 1917–1932.

Xue, F., H. Wang, and J. He, 2004: Interannual variability of Mascarene high and Australian high and their influences on East Asian summer monsoon. *J. Meteor. Soc. Japan*, 82, 1173–1186.

Yamada, H., K. Yoneyama, M. Katsumata, and R. Shirooka, 2010: Observations of a super cloud cluster accompanied by synoptic-scale eastward-propagating precipitating systems over the Indian Ocean. *J. Atmos. Sci.*, 67, 1456–1473.

Yasunaga, K., K. Yoneyama, Q. Moteki, M. Fujita, Y. N. Takayabu, J. Suzuki, T. Ushiyama, and B. Mapes, 2010: Characteristics of 3-4- and 6-8-day period disturbances observed over the tropical Indian Ocean. *Mon. Wea. Rev.*, 138, 4158–4174.

Yoneyama, K., Y. Masumoto, Y. Kuroda, M. Katsumata, K. Mizuno, Y. N. Takayabu, M. Yoshizaki, A. Shareef, Y. Fujiyoshi, M. J. McPhaden, V. S. N. Murty, R. Shirooka, K. Yasunaga, H. Yamada, N. Sato, T. Ushiyama, Q. Moteki, A. Seiki, M. Fujita, K. Ando, H. Hase, I. Ueki, T. Horii, C. Yokoyama, and T. Miyakawa, 2008: MISMO field experiment in the equatorial Indian Ocean. *Bull. Amer. Meteor. Soc.*, 89, 1889–1903.

Yoneyama, K., C. Zhang, and C. N. Long, 2013: Tracking pulses of the Madden-Julian oscillation. *Bull. Amer. Meteor. Soc.*, 94, 1871–1891.

Zhang, C., 2005: Madden-Julian Oscillation. *Rev. Geophys.*, 43, RG2003, doi:10.1029/2004RG000158.

Zhang, C., 2013: Madden-Julian oscillation bridging weather and climate. *Bull. Amer. Meteor. Soc.*, 94, 1849–1870.

Zhang, C., and M. Dong, 2004: Seasonality in the Madden-Julian oscillation. *J. Climate*, 17, 3169–3180.

Zuluaga, M. D., and R. A. Houze, 2013: Evolution of the population of precipitating convective systems over the equatorial Indian Ocean in active phases of the Madden–Julian oscillation. *J. Atmos. Sci.*, 70, 2713–2725.