The Onset of the Indonesian–Australian Summer Monsoon Triggered by the First-Branch Eastward-Propagating Madden–Julian Oscillation

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ABSTRACT

The Madden–Julian oscillation (MJO) often causes the onset of the Indonesian–Australian summer monsoon (IASM) over Indonesia and northern Australia. In the present study, a composite analysis is conducted to reveal the detailed IASM onset process and its air–sea interactions associated with the first-branch eastward-propagating MJO (FEMJO) based on 30-yr ERA-Interim data, satellite-derived sea surface temperature (SST), outgoing longwave radiation (OLR), and SODA3 ocean reanalysis. The results distinctly illustrate the phase-locked relationships among the persistent sea surface warming north of Australia, the FEMJO, and the established westerlies. It is found that the SST to the north of Australia reaches its annual maximum just before the onset of the summer monsoon. The oceanic surface mixed layer heat budget discloses that this rapid warming is primarily produced by the enhanced surface heat flux. In addition, this premonsoon sea surface warming increases the air specific humidity in the low-level troposphere and then establishes zonal moisture asymmetry relative to the FEMJO convection. This creates a more unstable atmospheric stratification southeast of the FEMJO and favors convection throughout the vicinity of northern Australia, which ultimately triggers the onset of the IASM. The results in this study thus may potentially be applicable to seasonal monsoon climate monitoring and prediction.

1. Introduction

The westerly Indonesian–Australian summer monsoon (IASM), usually beginning in December and ending in March (Figs. 1 and 2), brings heavy rainfall over the region of Indonesia and northern Australia (e.g., Troup 1961; Nicholls et al. 1982, 1984; McBride 1987; Manton and McBride 1992; Suppiah 1992; Drosowsky

1996; Chang et al. 2004, 2005; Wheeler and McBride 2005; Wang and Ding 2008; Zhang and Wang 2008; Colman et al. 2011; Robertson et al. 2011). Hence, the onset of the IASM indicates not only the abrupt reversal of wind fields but also the seasonal dry–wet transition in contrasting seasons. Naturally, the IASM has a governing climatic influence on international trade, local fisheries, agriculture, and even the social lifestyle of people in the monsoon region.

As shown in Figs. 1 and 2, the monthly characteristics show that the sea surface temperature (SST) in the monsoon region starts to increase steadily (approximately $0.03^{\circ}\text{C day}^{-1}$) 3 months before the monsoon

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FIG. 1. Monthly evolution of climatological SST (shaded; $^{\circ}\text{C}$), OLR (contours; only values lower than 230 W m^{-2} are plotted), and sea surface wind (vectors; m s^{-1}) from (a) October to (d) January. Cyan (black) vectors mean the surface zonal wind is eastward (westward). The blue box indicates the region of the Indonesian–Australian monsoon system in the present study.

transition. In addition, the zonal wind reversal generally coincides with the SST reaching its maximum in December. Moreover, the monthly outgoing longwave radiation (OLR) in the monsoon region also sharply decreases during October–December, and then the low-value branch (enhanced convection) veers southeastward to move across northern Australia. Although these remarkable changes occur during the pre-monsoon period, their monthly curves appear quite smooth (Fig. 2b), and particularly the drastic change in the propagating path of the convection and the sudden reversal of wind cannot be thus well demonstrated (Troup 1961), which implies that there must be some processes with time scales that are much shorter than the seasonal time scale to trigger the onset of the monsoon.

The mechanisms for the sudden seasonal changes in the IASM involves several different air–sea interaction processes as proposed and reviewed in previous works (e.g., Radok and Grant 1957; Troup 1961; Davidson et al. 1983, 1984; McBride 1983; Hendon et al. 1989; Joseph et al. 1991; Manton and McBride 1992; Suppiah 1992; Danielsen 1993; Hung and Yanai 2004; Wheeler and McBride 2005), and the convection-enhancing Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972) is thought to be a major trigger for the onset

of the IASM (e.g., McBride 1987; Lau and Chan 1988; Hendon and Liebmann 1990a,b; Hashiguchi et al. 1995; Kawamura et al. 2002; Hung and Yanai 2004; Wheeler and Hendon 2004). In the present work, we carry out the study from this hypothesis, since this mechanism is more consistent with the fact that the IASM onset presents a distinct southeastward-propagating pattern (Tanaka 1994; Zhang and Wang 2008). In fact, the same particular phenomenon in which the onset of the summer monsoon always coincides with the arrival of the convectively active intraseasonal oscillation (ISO) has also been noted in some other regional monsoon systems, such as the Indian summer monsoon (e.g., Joseph and Pillai 1988; Ghanekar et al. 2010; Zhou and Murtugudde 2014; Bhatla et al. 2017; Taraphdar et al. 2018), the Bay of Bengal summer monsoon (Li et al. 2013), and the South China Sea summer monsoon (e.g., Zhou and Chan 2005; Straub et al. 2006; Tong et al. 2009; Wu 2010; Kajikawa and Wang 2012; Lin et al. 2016).

Therefore, in the present study, first we use daily data to reexamine the detailed transitional process of the IASM from preonset to postonset regimes. The first-branch eastward-propagating MJOs (FEMJOs) are highlighted by magenta ellipses in each panel of Figs. 3 and 4. The year-by-year results clearly show that the strong and persistent SST warming is generally prior to

FIG. 2. (top) Time–latitude diagrams of monthly climatological SST (shaded; $^{\circ}\text{C}$), OLR (contours; W m^{-2}), and sea surface wind (vectors; m s^{-1}) averaged over 110° – 150°E . Cyan (black) vectors mean the surface zonal wind is eastward (westward). (bottom) Monthly evolution of the area-averaged SST (red line; $^{\circ}\text{C}$), OLR (blue line; W m^{-2}), and sea surface zonal wind (black line; m s^{-1}) in the monsoon region.

the onset of the monsoon and the zonal wind reversal always coincides with the arrival of the FEMJOs, which can be traced back to the tropical Indian Ocean (TIO) for several days. Prior to that, the easterlies are pronounced, and no MJOs pass through the Indonesian–north Australian region. This is consistent with the strong seasonality of the propagating path of the ISO in the eastern TIO (Madden and Julian 1971, 1972, 1994; Madden 1986; Wang and Rui 1990; Li and Wang 1994; Salby and Hendon 1994; Sperber et al. 2004; Zhang and Dong 2004; Li 2010; Li 2014; Wei and Hsu 2017). During austral winter, the maximum ISO activity is confined north of the equator, with pronounced northward or northeastward propagation over the eastern Indian Ocean (e.g., Yasunari 1979; Lawrence and Webster 2002; Jiang et al. 2004; Jiang and Li 2005; Li et al. 2013). During austral summer, the maximum ISO activity shifts to south of the equator, with predominant eastward or southeastward propagation (e.g., Wang and Rui 1990; Salby and Hendon 1994; Maloney and Hartmann 1998; Sperber et al. 2004; Kiladis et al. 2005; Zhang 2005; Hsu and Li 2012; Li 2014; Kim et al. 2017; Zhang and Ling 2017).

Recently, the onset of the Asian monsoon over the Bay of Bengal was clearly shown to be phase-locked to the local SST maximum and the arrival of the first-branch northward-propagating ISO (Li et al. 2013, 2016). In the present study, we attempt to investigate the onset process

of the IASM and its air–sea interaction associated with the FEMJO. Our main objectives of this study are 1) to identify the triggering role of the FEMJO in the onset of the IASM, 2) to examine the dynamic and thermodynamic processes that lead to the preonset SST warming, and 3) to explain the propagation mechanisms of the FEMJO associated with the preonset SST warming in the transitional season.

2. Data and methods

a. Data

The primary dataset used in the present study is the daily averaged ERA-Interim archive (Dee et al. 2011) from the European Centre for Medium-Range Weather Forecasts (ECMWF). The data analysis period is from 1981 to 2010, with a horizontal resolution of $1^{\circ} \times 1^{\circ}$ and 23 vertical pressure levels from 1000 to 200 hPa. The three-dimensional variables used for analysis include zonal and vertical wind components, divergence, temperature, and specific humidity, and the two-dimensional variables used are the 10-m wind, heat flux, and cloud-cover fields.

The gridded daily $1^{\circ} \times 1^{\circ}$ OLR–Climate Data Record data from NOAA’s National Climatic Data Center (Lee 2014) are used as a main proxy for convective activities. The daily NOAA high-resolution blended OISST V2

FIG. 3. Longitude–time sections of daily SST (shaded; $^{\circ}\text{C}$), sea surface wind (vectors; m s^{-1}), and 20–90-day bandpass-filtered OLR anomalies (contours; W m^{-2} ; only negative values are plotted) averaged between 15°S and the equator for 1981–95. The FEMJO, which originates in the southwestern Indian Ocean and first propagates through the monsoon region every year, is highlighted with magenta ellipses.

FIG. 4. As in Fig. 3, but for 1996–2010.

data (Reynolds et al. 2007) provided by the remote sensing system with a resolution of $0.25^\circ \times 0.25^\circ$ are used to investigate the air–sea interaction processes associated with the FEMJO.

To examine the control mechanisms of the pre-monsoon SST warming, the SODA, version 3.4.2, 5-day output is used to diagnose the mixed layer heat budget. This dataset, with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ and 17 vertical levels in the upper 200-m layer, is also forced with the ERA-Interim fields (Carton et al. 2018).

For all dynamic and thermodynamic variables in the present result analyses, unless specified otherwise, all the data are bandpass filtered with a 20–90-day Lanczos digital filter (Duchon 1979) to obtain their intraseasonal components. To illustrate the role of the background conditions in causing the southeastward propagation of FEMJO convection, the low-frequency component with a period longer than 90 days is extracted by a low-pass filter. In addition, all the climatology fields are derived based on the long-term means for 1981–2010.

b. Methods

1) DEFINITION OF THE MONSOON ONSET DATE

Traditionally, the onset of the IASM has been defined using the low-level westerly wind, precipitation, large-scale circulation, and cloud criteria (e.g., Troup 1961; Nicholls et al. 1982; Davidson et al. 1983; Holland 1986; Hendon and Liebmann 1990a; Tanaka 1994; Drosowsky 1996; Hung and Yanai 2004). In the present study, in order to identify the triggering role of the FEMJO in the onset of the IASM, the convective activity indicated by the intraseasonal OLR over the monsoon region (0° – 15° S, 110° – 150° E; see blue box in Fig. 1) is used solely to determine the onset date. The first day when the maximum FEMJO convection, indicated with the lowest OLR anomalies, arrives at 120° E is chosen as the onset day (Fig. 5). The 30-event mean onset date in the present study is 11 December, which is slightly earlier than the previous results obtained by using other criteria [24 December by Holland (1986); 25 December by Hendon and Liebmann (1990a); 22–26 December by Tanaka (1994); 28–29 December by Drosowsky (1996); 25 December by Hung and Yanai (2004)]. This is because the center of the active FEMJO convection observed from the OLR generally precedes the westerly wind bursts and MJO precipitation by several days (Madden and Julian 1972; Tanaka 1994; Hung and Yanai 2004; Wang et al. 2018). The standard deviation of the onset date obtained in the present study is 15 days, which is also close to the previous results [15 days by Holland (1986); 16 days by Hendon and Liebmann (1990a); 14 days by Hung and Yanai (2004)].

To further confirm the selection of the monsoon onset date, following the method of Wheeler and McBride (2005), the relationship between the local OLR-defined monsoon onset in this study and the state of the globally defined Real-time Multivariate MJO(RMM) index defined by the leading pair of EOFs of equatorial averaged zonal wind and OLR (Wheeler and Hendon 2004) is presented in Fig. 6. Considering only the dates that lie outside the central unit circle (i.e., those occurring when the MJO is nonweak and can be discerned using the RMM methods), the figure clearly displays that the onset occurs more than 90% of the time when the FEMJO is in phases 4–6. Therefore, the low-level westerlies and broad-scale convection of the FEMJO are in the vicinity of northern Australia. The composited spread of the onsets from phases 4 to 6 covers a time window of approximately 21 days, as shown in Fig. 6. These patterns are consistent with the previous results (McBride 1983; Tanaka 1994; Wheeler and Hendon 2004; Wheeler and McBride 2005; Zhang and Wang 2008).

2) SURFACE MIXED LAYER HEAT BUDGET

The simplified mixed layer heat balance equation as presented by Foltz et al. (2010) can be written as

$$\rho c_p h \frac{\partial T_{\text{mld}}}{\partial t} = Q_0 - \rho c_p h U \cdot \nabla T_{\text{mld}} + R, \quad (1)$$

where h is the mixed layer depth (MLD), T_{mld} is the average surface mixed layer temperature (MLT), U is the average surface mixed layer currents, c_p is the heat capacity, and ρ is the seawater density. The terms in Eq. (1) represent, from left to right, the surface mixed layer heat storage rate; the net surface heat flux (corrected for the penetration of shortwave radiation through the base of the surface mixed layer); the horizontal advective heat flux; and the residual flux including the horizontal divergence of the eddy heat flux within the mixed layer, the entrainment and vertical turbulent heat flux across the base of the surface mixed layer, and the analysis and sampling errors in the estimation of the other terms in Eq. (1). The reason we just use the simplified MLT budget will be discussed in section 4.

The net surface heat flux Q_0 is the sum of the latent heat flux, sensible heat flux, longwave and shortwave radiation, and penetrative shortwave radiation through the base of the surface mixed layer:

$$Q_0 = Q_{\text{LH}} + Q_{\text{SH}} + Q_{\text{LW}} + Q_{\text{SW}}[1 - f(h)], \quad (2)$$

$$f(h) = R e^{-h/l_1} + (1 - R) e^{-h/l_2}. \quad (3)$$

We use $R = 0.62$, $l_1 = 0.6$ m, and $l_2 = 20$ m, which are coefficients that depend on water turbidity, as classified by Jerlov (1968).

FIG. 5. (a) Time–latitude sections of the 20–90-day-filtered OLR anomalies at 120°E (only negative values are shaded, and contour interval is 5 W m^{-2}). The zones between 15°S and the equator are shown from 1 Nov to 31 Jan for each season from 1981/82 to 2010/11. The thick green lines denote the IASM onset days in this study, as determined by the FEMJO convection center (the lowest OLR anomalies) reaching 120°E . (b) Time–latitude composite of the 20–90-day-filtered OLR for 60 days before and after the monsoon onset.

FIG. 6. Onset date each year as a function of the state of the RMM index. The gray two-digit numbers plotted refer to the monsoon year. The black dot indicates the composited onset date, and the blue and red lines illustrate the progression of the composited FEMJO.

3) ATMOSPHERIC CONVECTIVE INSTABILITY

Atmospheric convective instability is a main background factor that possibly affects MJO propagation. The convective instability parameter is defined as the difference in the equivalent potential temperature θ_e between the lower and middle troposphere (Zhang et al. 2004; Ding and He 2006):

$$\Delta\theta_e = \theta_e|_{1000-700 \text{ hPa}} - \theta_e|_{600-300 \text{ hPa}}. \quad (4)$$

A positive (negative) value of $\Delta\theta_e$ implies that the atmosphere is potentially unstable (stable). The $\Delta\theta_e$ pattern could inform us where the atmospheric conditions favor the development of convection.

3. Composite evolution features of the FEMJO and monsoon onset

Based on the above monsoon onset date derived from the 30 FEMJO cases (Fig. 5), we can easily obtain a composite map of the summer monsoon onset process (Fig. 7) and the time sequence (with a 5-day interval) of the composite evolution patterns (Fig. 8). It is obvious that distinct sea surface warming occurs 3 months before the reversal of the wind directions, and forms a pronounced hot spot to the north of Australia. This SST reaches its annual maximum value approximately

FIG. 7. Composite longitude–time sections averaged between 15°S and the equator: SST (shaded; °C), sea surface wind (vectors; m s^{-1}), and 20–90-day bandpass-filtered OLR anomalies (contours; W m^{-2} ; only negative values are plotted). Day 0 is a reference day when maximum FEMJO convection arrives at 120°E, which is indicated by the 20–90-day bandpass-filtered OLR with a Lanczos digital filter.

10 days before the arrival of the FEMJO, with a value of $>30^\circ\text{C}$ at approximately $125^\circ\text{--}127^\circ\text{E}$. In addition, the sustained westerlies are established, which is concurrent with the first successive eastward movement of the convection event entering into the monsoon region. Before that, convection is mainly confined to the north of the equator, and the easterlies are pronounced in the monsoon region. It should also be noted that the wind speed is relatively weak during the premonsoon period.

Note that the Maritime Continent is a known barrier for the eastward propagation of the MJO (Nitta et al. 1992; Zhang and Ling 2017; Kim et al. 2017). Figure 8 displays that the composite FEMJO convection initiates in the TIO at day -20 . Then, it quickly moves southeastward, with a magnitude of approximately 5.1 m s^{-1} . At day -10 , the major convection arrives at the Maritime Continent. Subsequently, the major convective branch slowly shifts farther southeastward over northern Australia, where it becomes much stronger. At day 0, the OLR anomaly has reached its most southerly extent centered at 12°S . At the same time, the westerly wind begins to dominate the whole monsoon region.

When the enhanced FEMJO travels slowly for approximately 15 days over the northern Australian longitudes, the largest signal of the convection in the OLR appears clearly over the warmer sea (Fig. 8). Many previous studies have stated that complex air–sea interactions play an essential role in maintaining local, stationary ISOs in deep convection (e.g., Manton and

FIG. 8. Composite evolution patterns of the SST (shaded; $^{\circ}\text{C}$), sea surface wind (vectors; m s^{-1}), and 20–90-day-filtered OLR anomalies (black contours; W m^{-2} ; only negative values are plotted).

McBride 1992; Hendon and Liebmann 1990a; Hirst and Lau 1990; Hu and Randall 1994; Wang and Xie 1998; Watterson and Syktus 2007; Lin et al. 2011; Hsu and Li 2012). Therefore, the increasing SST is hypothesized to precondition the onset of the IASM through its role in steering the MJO from its boreal summer state to its austral summer mode. Nevertheless, the control mechanism for the premonsoon SST warming and its role in inducing the FEMJOs has not been clearly addressed. More generally, it is not clear why the TIO MJO veers eastward from its previous northward propagation route

during the boreal autumn. The present study was thus conducted to explore the potential impacts of the oceanic processes. Specifically, the aim was to identify the role of the premonsoon SST in driving the FEMJO and, hence, initiating the summer monsoon.

4. Mechanisms of the premonsoon SST warming

The persistent SST warming in northern Australia begins approximately 3 months before the onset of the IASM, during which the maximum positive tendency

FIG. 9. (a) Composite longitude–time sections averaged between 15°S and the equator: daily SST (shaded; °C), net surface heat flux (black contours; W m^{-2}), and the OLR anomalies (green contours; W m^{-2}) associated with FEMJO as in Fig. 7. (b) As in (a), but for SST (shaded; °C) and MLT (contours; °C) from SODA products. Also shown is the MLD heat budget (c) along 120°–140°E and (d) for the premonsoon period (day –90 to day –10).

reaches $0.05^{\circ}\text{C day}^{-1}$, and the total increment is nearly 4.0°C in the 90 days (Fig. 9a). Both of these daily results are more significant than those obtained from the climatology monthly data (Fig. 2). To clarify the controlling mechanisms of this premonsoon SST warming, we diagnosed the oceanic surface mixed layer heat budget by using the pentad-averaged SODA outputs. Since the SODA assimilated the satellite SST and a large amount of in situ upper-ocean temperature and salinity profiles, it reproduced the observations well (see Figs. 9a,b).

The mixed layer heat budget analysis (Figs. 9c,d) displays clearly that the mixed layer heat change rate is positive and stable at approximately $18.9 \pm 4.5 \text{ W m}^{-2}$ during the premonsoon period. Among the three main factors, the net surface heat flux ($40.5 \pm 7.2 \text{ W m}^{-2}$) and the residual flux ($-22.1 \pm 2.6 \text{ W m}^{-2}$) are dominant in this heat change, and the much smaller zonal advection component ($<1.0 \text{ W m}^{-2}$) can be neglected. The results obtained here are quite consistent with the modeling study by Santos et al. (2010). The net surface heat flux undoubtedly plays a primary role in the premonsoon sea surface warming. The only caveat to this result is the large residual in the heat budget. In fact, during the preonset regime, the southeastward wind dominates the monsoon region, and these alongshore winds induce some strong local coastal upwelling systems in this region, as described by previous studies (e.g., Wyrtki 1962;

Susanto et al. 2001; Du et al. 2005; Qu et al. 2005; Siswanto and Suratno 2008; Chen et al. 2016; Ningsih et al. 2013). Additionally, the oceanic stratification in the tropical region with cooler water underlying warmer water is very stable, and the vertical heat exchange through the base of the mixed layer could only have a negative impact on the above SST warming.

The sun moves southward from October to December, and the corresponding enhanced solar radiation gradually warms the Southern Hemisphere. However, why is this significant SST warming phenomenon just confined to northern Australia and the surrounding seas, while the warming of the other ocean regions at the same latitudes are much slower (Figs. 7, 8)? As shown in Fig. 9a, the net heat flux that the ocean gains from the atmosphere in the monsoon region (110° – 150°E) generally exceeds 100 W m^{-2} during the premonsoon period, while the other regions at the same latitudes are only 40 W m^{-2} . This large difference is interrupted by the intraseasonal perturbation associated with the FEMJO. Considering the four components separately, as shown in Fig. 10, the large difference in the zonal distribution of net surface heat flux during the premonsoon period is mainly due to more shortwave radiation being absorbed by the ocean and the reduced loss by latent heat flux resulting from the weaker wind speed as mentioned above in the monsoon region (Hendon

FIG. 10. Composite longitude–time sections of surface heat flux component (W m^{-2}) averaged between 15°S and the equator: (a) shortwave radiation, (b) latent heat flux, (c) longwave radiation, and (d) sensible heat flux. Positive heat flux shows a gain to the ocean. The black contour in (a) indicates the corresponding total cloud cover and in (b) is the wind speed.

et al. 2012; see also Fig. 7). At the onset of the IASM, the increased cloudiness and rainfall result in decreasing shortwave radiation, while stronger winds induce an increase in the loss of latent heat flux from the ocean to the atmosphere.

5. Role of the premonsoon SST warming in driving the FEMJO

The composite evolution patterns in Figs. 7 and 8 illustrate that FEMJO convection starts to propagate southeastward after passing the central TIO. A natural

question is why the FEMJO veers eastward from its previous northward propagation route in the eastern TIO. To address this question, we need to examine the structure of the FEMJO in the first instance. Here, the zonal–vertical structure (averaged between 15°S and the equator) of the composite FEMJO derived from ERA-Interim was made relative to the convection center reaching 120°E (Fig. 11). It displays that a marked zonal asymmetry appears in the low-level specific humidity and θ_e fields (Figs. 11d,e), with a notable positive anomaly appearing to the east of the MJO convection. Additionally, a maximum perturbation convergence in

FIG. 11. Zonal–vertical distributions of the composite fields when the FEMJO center reaches at 120°E : (a) zonal wind (m s^{-1}), (b) vertical velocity (m s^{-1}), (c) divergence (10^{-6} s^{-1}), (d) specific humidity (g kg^{-1}), and (e) θ_e (K) averaged between 15°S and the equator. Solid (dashed) lines indicate positive (negative) values.

FIG. 12. (a) Spatial distribution of the background convective instability field (low-frequency component with a period longer than 90 days; K) during day -20 to day 0 . (b) Longitude–time section of the background convective instability field (shaded; K) and the OLR anomalies averaged between 15°S and the equator associated with FEISO (contours; W m^{-2}). Solid (dashed) lines indicate positive (negative) values.

the PBL (1000–700 hPa) appears at approximately 1000 km east of the convection center (Fig. 11c). According to previous studies (e.g., Seo and Kim 2003; Maloney 2009; Hsu and Li 2012; Kim et al. 2013; Hsu et al. 2014; Li 2014; Jiang 2017; Jiang et al. 2018), such PBL moistening and convergence distribution would precondition the convective instability to the east of the MJO convection and then lead to the eastward propagation of the FEMJO.

To confirm the above theories, we consequently examine the distribution of the background convective instability parameter $\Delta\theta_e$ averaged between day -20 and day 0 (Fig. 12a). It displays that the maximum background convective instability in the TIO appears within 10° of the equator, and a significant increase in the $\Delta\theta_e$ appears south of 10°S , northwest of Australia, consistently with the above PBL moistening to the east of the FEMJO convection. This result means the low-frequency atmospheric background state is potentially more unstable to the southeast of the FEMJO convective center. Therefore, a phase leading to a positive low-level moisture anomaly may form a relatively unstable stratification and generate a favorable environment for the potential development of new convection to the

southeast of the FEMJO convection center, which is consistent with the FEMJO convection behavior in the eastern TIO (see Fig. 8). Figure 12b also illustrates the temporal evolution of the background convective instability field averaged between 15°S and the equator. Note that the $\Delta\theta_e$ values are negative east of 140°E prior to day -60 . During that time, the strong ISO activity is mainly confined in the equatorial region and moves northward in the eastern TIO. Subsequently, $\Delta\theta_e$ gradually increases in the Southern Hemisphere, and at approximately days -20 to 20 , the average value over northern Australia is much greater than that in the Northern Hemisphere. This basic change in the atmospheric conditions veers the path of the FEMJO. Next, we address how the asymmetric background condition in the eastern TIO is established.

To quantitatively measure the respective contributions of this convective instability from the variation in air temperature T_a and specific humidity q , we follow the method of Li et al. (2013) and define a convective instability parameter as the average of $\Delta\theta_e$ from 120° to 140°E and from 15°S to the equator, where the largest background convective instability occurs (see Fig. 12). The parameter exhibits a continuously increasing trend during the premonsoon period and reaches its maximum at day -10 (black line in Fig. 13a). Then, we recalculate $\Delta\theta_e$ using either (T_a, q_0) or (T_0, q) , with (T_0, q_0) being the value at day -90 . The result clearly shows that the increase in $\Delta\theta_e$ is primarily attributed to the variation in the specific humidity field (red line in Fig. 13a). Considering the premonsoon period from day -60 to day 0 , we next identify the relative contributions of the specific humidity at the low level (1000–700 hPa) and upper level (600–300 hPa) to the $\Delta\theta_e$ increase. Figure 13b shows that $\Delta\theta_e$ increases from day -90 to day 0 by 5.9 K and this increase is mainly contributed by low-level $\Delta\theta_e$ (9.7 K), while the upper-level $\Delta\theta_e$ has a negative contribution of -3.8 K . Thus, in general, the enhanced background convective instability over northern Australia is primarily attributed to the increase in the low-level q .

The diagnosis above reveals that the low-level moisture-induced atmospheric background state plays the dominant role in causing the eastward propagation of the FEMJO in the eastern TIO. In fact, enhanced convective instability is induced in the monsoon region due to the significant increase in the surface θ_e values, and the θ_e values over northern Australia during the premonsoon period are generally in a transitional mode between the winter and summer (Fig. 14). All of these results imply that the surface processes associated with the ocean play a leading role in the seasonal evolution of convective instability and then triggers the IASM by

FIG. 13. (a) Time evolution of $\Delta\theta_e$ (black line; K) and the partial contributions of $\Delta\theta_e$ due to the temperature (blue) and moisture (red) changes averaged over the region 15°S – 0° , 120° – 140°E . (b) Difference (day 0 minus day –90) of the background convective instability parameter (defined as an averaged value of $\Delta\theta_e$ over 15°S – 0° , 20° – 140°E ; black bar; K) and relative contributions from the low level (averaged over 1000–700 hPa; red bar; K) and the upper level (averaged over 600–300 hPa; blue bar; K).

driving the FEMJO. The substantial sea surface warming to the north of Australia during premonsoon period could impact the θ_e values by modulating the local surface T_a and q . Therefore, considering the SST– T_a , SST– q , T_a – q , and T_a – r relationships in this paper, there are distinct contrasts with regard to the onset of the summer monsoon (Fig. 15). During the calm premonsoon period, T_a (q) increases linearly by approximately 0.7°C (0.7 g kg^{-1}) with the underlying SST (see Figs. 15a,c). However, the relative humidity r generally remains constant (approximately 77%) in the calm premonsoon period (Fig. 15d). At the onset of the summer monsoon, the above relationships significantly change and the parameters do not correlate well with SST. In fact, according to the Clausius–Clapeyron relationship, q increases with the warmer air, which is able to hold more water vapor while r remains almost constant. That means the saturation vapor pressure would increase, theoretically, by nearly 6% per 1°C of warming

at the reference temperature of 27.0°C (Fig. 15b). In our results, the observed rate of q increases with T_a is $5.7\% ^\circ\text{C}^{-1}$, which is quite close to the theoretical value.

6. Conclusions and discussion

In this study, the detailed onset process of the IASM and related air–sea interactions over the Indonesia–northern Australia region are investigated through a composite diagnosis of ERA-Interim, OLR, and SST data for the period of 1981–2010. The main results are listed below:

- 1) The onset of the IASM is phase-locked to the local seasonal SST maximum and the arrival of the FEMJO originating from the TIO.
- 2) During the premonsoon period, enhanced net sea surface heat flux, which is primarily attributed to the enhanced solar heating, in conjunction with the weak

FIG. 14. Zonal-vertical distributions of θ_e (K) in the low-level troposphere averaged between 15°S and the equator in (a) JJA, (b) DJF, and (c) the premonsoon period.

FIG. 15. Composite relationships, derived from ERA-Interim in the region 15°S – 0° , 120° – 140°E between (a) the SST and air surface temperature, (b) air surface temperature and specific humidity, (c) SST and surface specific humidity, and (d) air surface temperature and relative humidity. The blue circles indicate the period of day 0 to day +60, and the asterisks represent the period of day –60 to day 0, with the premonsoon period (day –40 to day –10) being denoted in red. The linear regressions for the premonsoon period are given in (a)–(c).

wind conditions leads to a distinct sea surface warming to the north of Australia, and makes the SST reach its annual maximum just before the onset of the summer monsoon.

- 3) The premonsoon SST warming increases the low-level specific humidity of the air and then establishes zonal moisture asymmetry relative to the FEMJO convection. These pre-established background conditions precondition the atmospheric convective instability to the southeast of the FEMJO convection center, leading to the onset of new areas of convection and thus the southeastward propagation, which ultimately triggers the IASM.

We expand the results of [Hendon and Liebmann \(1990a\)](#) and [Kawamura et al. \(2002\)](#), and relate the onset

of the IASM with the FEMJO. As suggested in the present study, understanding and predicting the FEMJO thus will be crucial for predicting the onset of the IASM and for agricultural planning and water management in Indonesia and northern Australia.

It should be pointed out that the arrival of the FEMJO triggers the onset of the IASM for most cases. By taking a global view of this regional monsoon system, there is one onset case in the present study (see [Fig. 6](#)) that occurs when the FEMJO is relatively weak (i.e., when it is difficult to discern using the RMM methods) and there are two cases associated with phases 4 and 6 (i.e., when northern Australia is in the suppressed phase of the MJO). As discussed by [Hendon and Liebmann \(1990a\)](#), when using a local definition of the MJO as the present paper did (20–90-day bandpass-filtered local

OLR anomalies), the obtained MJO will limit the onset of the monsoon to within its active phase, but the actual onset may be set by other synoptic phenomena as proposed by the previous studies (e.g., Davidson et al. 1983; Hendon et al. 1989; Danielsen 1993; Drosowsky 1996; Hung and Yanai 2004; Wheeler and McBride 2005). Despite all the above limitations, the results in the present study still provide us a more effective application for local seasonal monsoon climate monitoring and prediction.

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