Structure and Origin of the Quasi-Biweekly Oscillation over the Tropical Indian Ocean in Boreal Spring

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ABSTRACT

The structure and evolution features of the quasi-biweekly (10–20 day) oscillation (QBWO) in boreal spring over the tropical Indian Ocean (IO) are investigated using 27-yr daily outgoing longwave radiation (OLR) and the National Centers for Environment Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data. It is found that a convective disturbance is initiated over the western IO and moves slowly eastward. After passing the central IO, it abruptly jumps into the eastern IO. Meanwhile, the preexisting suppressed convective anomaly in the eastern IO moves poleward in the form of double-cell Rossby gyres. The analysis of vertical circulation shows that a few days prior to the onset of local convection in the eastern equatorial IO an ascending motion appears in the boundary layer.

Based on the diagnosis of the zonal momentum equation, a possible boundary layer–triggering mechanism over the eastern equatorial IO is proposed. The cause of the boundary layer convergence and vertical motion is attributed to the free-atmospheric divergence in association with the development of the barotropic wind. It is the downward transport of the background mean easterly momentum by perturbation vertical motion during the suppressed convective phase of the QBWO that leads to the generation of a barotropic easterly—the latter of which further causes the free-atmospheric divergence and, thus, the boundary layer convergence. The result suggests that the local process, rather than the eastward propagation of the disturbance from the western IO, is essential for the phase transition of the QBWO convection over the eastern equatorial IO.

1. Introduction

Atmospheric subseasonal oscillations contain two well-separated bands with 10–20- and 30–60-day periods, respectively (Chen and Chen 1993; Fukutomi and Yasunari 1999; Annamalai and Slingo 2001; Li and Wang 2005; Goswami 2005; Li and Zhou 2009). The 30–60-day oscillation was first discovered by Madden and Julian (1971, 1972). A major feature of the Madden–Julian oscillation (MJO) is the eastward propagation along the equator. The later studies further documented that in addition to the eastward propagation, the tropical intraseasonal (30–60 day) oscillation (ISO) also has a distinctive northward propagation in the monsoon region (Yasunari 1979) and a westward propagation off the equatorial region (Wang and Rui 1990) in boreal summer. In the present work, we use the term MJO to represent the general 30–60-day mode of ISO over the tropical region.

The 10–20-day oscillation in the tropics was reported by several previous studies (e.g., Wallace and Chang 1969; Nitta 1970; Chang et al. 1970; Yanai and Murakami 1970; Murakami 1971). It was referred to as the quasi-biweekly oscillation (QBWO) by Murakami (1976) and Krishnamurti and Bhalme (1976), who pointed out the remarkable QBWOs that exist in the entire monsoon...
region. This oscillation was found to be closely related to the active and break phases of the summer monsoon (Yasunari 1979; Krishnamurti and Ardanuy 1980; Chen and Chen 1993; Zhou and Chan 2005, etc.).

While the MJO exhibits a zonal wavenumber-1 structure, the QBWO is generally around zonal wavenumber 6 in boreal summer or winter (Kiladis and Wheeler 1995; Annamalai and Slingo 2001). For the vertical structure, the MJO has the first baroclinic mode structure, with a node at about 500 hPa. In contrast, the QBWO exhibits an equivalent barotropic structure, extending from the lower to the upper troposphere with a reversed phase near 100 hPa (Chen and Chen 1993; Kiladis and Wheeler 1995; Annamalai and Slingo 2001; Chatterjee and Goswami 2004; Fukutomi and Yasunari 2005; Yokoi and Satomura 2006). Except for the northern spring, the QBWO is found to have maximum variances off the equator and propagate westward. For example, in boreal summer the QBWO convective activity may propagate all the way from the western tropical Pacific to the Indian Peninsula, to affect the active and break of the Indian summer monsoon (Murakami 1980; Annamalai and Slingo 2001; Chen et al. 2004; Lin and Li 2008). The phase difference between the boundary layer convergence and the convection may be responsible for the westward propagation of the QBWO (Goswami 2005).

So far the understanding of the cause of the QBWO is limited. Krishnamurti and Bhalme (1976) proposed that the cloud–radiation–convection feedback may maintain the QBWO over the Indian monsoon region. Goswami and Mathew (1994) and Chatterjee and Goswami (2004) suggested that the QBWO is an intrinsic mode of the tropical atmosphere driven by evaporation–wind feedback in the presence of the mean westerly or boundary layer convergence–induced convective heating. Chen and Chen (1993) noted that low-level circulation associated with the QBWO in boreal summer exhibits a double cyclonic–anticyclonic cell structure, with the northern structure centered at about 15°–20°N and the southern one at the equator. This equatorially asymmetric structure is possibly attributed to the influence of the background summer mean flow (Chatterjee and Goswami 2004). Kiladis et al. (1994) and Kiladis and Wheeler (1995) noted that equatorially symmetric Rossby waves in the period of 6–30 days appear over the tropical Pacific, and those waves move westward with eastward energy dispersion. Wen and Zhang (2008) detected Rossby wave–like circulation related to the QBWO over the eastern tropical Indian Ocean (IO) in boreal spring. They suggested that feedbacks among the convection, Rossby wave response, and associated low-level circulation are crucial in maintaining the observed QBWO structure.

Given the distinctive propagation characteristics of the QBWO in different seasons, it is anticipated that the origin of the QBWO might be different. In this study, we focus on the equatorially symmetric mode of the QBWO in boreal spring. The objective of the current study is to explore the 3D structure and propagation features of the QBWO over the tropical IO and to investigate the specific physical mechanism that gives rise to the QBWO over the tropical IO. The rest of this paper is organized as follows: In section 2, we describe datasets and methods used. The seasonal characteristics of the QBWO activity over the tropical IO are compared in section 3. The evolution of the 3D structure of the atmospheric circulation associated with the QBWO in boreal spring is examined in section 4, and a mechanism for the origin of the QBWO in the tropical IO is proposed in section 5. Finally, discussion and conclusions are given in section 6 and section 7, respectively.

2. Data and methodology

Two major datasets are used in this work. One is the daily reanalysis from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR; Kalnay et al. 1996), including zonal and meridional winds \( u \) and \( v \), geopotential height \( z \), specific humidity \( q \), and vertical velocity \( \omega \) at pressure levels. The other is the daily National Oceanic and Atmospheric Administration (NOAA) outgoing longwave radiation (OLR) data, used as a proxy for the tropical convection (Liebmann and Smith 1996). Both of the datasets have a global coverage with a 2.5° longitudinal and latitudinal resolution, spanning from 1979 to the present. [Datasets are provided by the NOAA’s Office of Oceanic and Atmospheric Research/Earth System Research Laboratory (NOAA/OAR/ESRL) Physical Sciences Division (PSD), Boulder, Colorado, available online at http://www.cdc.noaa.gov/]

Lanczos filtering is used on the daily data to get the subseasonal signals. As introduced by Duchon (1979), the Lanczos filtering is digital filtering that uses a linear relationship to transform the raw data sequence \( x_t \) into a filtered data sequence \( y_t \) as

\[
y_t = \sum_{k=-n}^{n} w_k x_{t+k},
\]

where the \( w_k \) are smoothed weights. For bandpass filtering, the smoothed weights are written as

\[
w_k = \frac{(\sin 2\pi f_c k / \pi k) - \sin 2\pi f_s k / \pi k)}{(\pi k / \pi k)} \sin \pi k/n, \quad k = -n, \ldots, 0, \ldots n.
\]
Here, two the cutoff frequencies \( f_{c2} \) and \( f_{c1} \) are 0.1 and 0.05 day\(^{-1}\) for QBWO, respectively, and 0.033 and 0.017 day\(^{-1}\) for MJO, respectively. The total number of weights \( n \) are chosen in such a way that the response functions are close to 1 between two cutoff frequencies and to 0 beyond (for QBWO, \( n = 44 \) and for MJO, \( n = 79 \)). Limited by the values of \( n \), the valid filtered data used in this analysis are from 1980 to 2006.

A lagged regression is applied to the 27-yr daily data to examine the evolution of the coherent 3D atmospheric circulation patterns associated with the QBWO convection. A \( t \) test is used to determine the statistical significance of the regressed fields. As for the effective sample size (ESS) for the \( t \) test, we adopt a method introduced by Bretherton et al. (1999), as Wen and Zhang (2008) also did. Considering the discontinuity of the daily time series (only 92 days from 1 March to 31 May are used each year), the ESS is first calculated individually for each year, and then the sum of them is regarded as the total ESS. For the current study, a time series of 27 \( \times \) 92 days corresponds to a mean ESS of 500.

3. Seasonal dependence of the QBWO over the tropical IO

Figure 1 shows the climatological mean of OLR and the QBWO and MJO variances for annual mean and four natural seasons, respectively. For the climatological annual mean field, the most vigorous convection appears over the eastern and central tropical IO, with a minimum OLR center over Sumatra. Strong convective activities associated with MJO and QBWO also appear in the region where the mean convection is pronounced. In general, the variance distributions of QBWO and MJO resemble each other over the IO, except that the QBWO variance center extends farther northward to the Bay of Bengal (BOB).

While the patterns of the mean OLR and the QBWO and MJO variances in boreal spring [March–April–May (MAM)] are similar to those of the annual mean, being symmetric about the equator, they are quite asymmetric in other seasons. For instance, in boreal summer [June–July–August (JJA)] the most vigorous convection shifts northward to the Indian Peninsula, BOB, and the Indochina Peninsula, while the maximum QBWO variances appear over the BOB and the eastern part of the Arabian Sea. In northern autumn [September–October–November (SON)], the mean convective belt shifts to the equatorial region, but the maximum variances of QBWO and MJO appear north of 10°N. In boreal winter [December–January–February (DJF)], both the mean convection and the maximum variance centers lie over the southern tropical IO.

For all seasons, the amplitude of the QBWO is similar to that of MJO. This indicates that the QBWO is equally important in contributing to the atmospheric variability in the tropics.

The empirical orthogonal function (EOF) analysis method is employed to reveal the dominant patterns of the QBWO over the tropical IO at each season. The structures of the first and second EOF modes (EOF1 and EOF2) of the 10–20-day-filtered OLR field and the time-lag correlation coefficients between the two principal components (PCs) are shown in Fig. 2. The positive (negative) lagged time denotes that PC2 lags (leads) PC1. Also, illustrated in Fig. 2 are the regressed 850-hPa wind fields. Note that in boreal spring both the first and second EOF modes represent an equatorially symmetric pattern over the tropical IO, which is distinctive from the other seasons.

In EOF1, an active convection (negative OLR) center appears east of 80°E and a suppressed convection (positive OLR) center appears in the western IO. The intensity of the former is much stronger than that of the latter. On the east edge of the negative OLR center, the contours are approximately parallel to the shoreline of Sumatra Island, implying a possible impact of Sumatra on local convective activities (Nitta et al. 1992). The associated 850-hPa winds over the eastern IO exhibit a double-cell cyclonic Rossby gyre structure symmetric about the equator, with anomalous westerlies being right over the equator. A weak anticyclonic Rossby gyre pattern may be identified over the western IO, in association with the suppressed convection there. The symmetric OLR and wind patterns resemble those shown in Kiladis and Wheeler (1995) and Wen and Zhang (2008).

In EOF2, a broadscale positive OLR anomaly occurs over the equatorial IO, with a maximum center around 80°E. Over the eastern IO, this positive OLR anomaly is accompanied by two negative centers at 15°S and 15°N, making up a “sandwich” pattern there. The negative OLR centers are associated with cyclonic Rossby gyre pairs at 850 hPa. An interesting feature is the in-phase relationship between the positive OLR and westerly anomalies over the equatorial eastern IO. The possible mechanism for such an in-phase relationship will be discussed in the following sections.

A significant lagged correlation between EOF1 and EOF2 suggests that physically they represent the different phases of the same QBWO cycle (Lau and Lau 1990; Plaut and Vautard 1994; Jiang et al. 2004; Jiang and Li 2005). Thus, one may use the time series of EOF1 to reveal the evolution characteristics of the entire QBWO cycle.

The EOF patterns in boreal summer show a more asymmetric structure, with the centers of negative and positive OLR anomalies shifting northward to 5°–10°N. Accordingly, the 850-hPa winds exhibit double Rossby
Fig. 1. (left) Climatological mean of OLR (W m\(^{-2}\)), (middle) QBWO (10–20 day) variance (W\(^2\) m\(^{-4}\)), and (right) MJO (30–60 day) variance (W\(^2\) m\(^{-4}\)) for (a) annual mean, (b) MAM, (c) JJA, (d) SON, and (e) DJF.
gyre cells, with the axis along $5^\circ$–$10^\circ$N. When the analysis domain is extended eastward to include the tropical western Pacific, the maximum activity center is found off the equator in the western Pacific. This is consistent with Chen and Chen (1993), who found that the QBWO OLR anomalies in boreal summer originate over the western tropical Pacific and move westward. Having the similar amount of variances, the EOF1 and EOF2
modes in boreal summer over the IO are not significantly correlated.

Similar to the wind patterns in summer, the axis of the double cells in northern fall is about 5°–10°N of the equator. The significant lagged correlation between the first and second EOF modes in autumn implies that the OLR anomalies move northeastward from the equatorial IO to the west coast of Indochina Peninsula.

In DJF, the major QBWO activity is confined at and south of the equator. It again exhibits an equatorial asymmetric feature. The lagged correlation between the two dominant EOF modes is not significant.

To summarize, the QBWO convection and low-level circulation exhibit distinct characteristics in the four seasons. Only in boreal spring are the patterns of both the convection and low-level circulation anomalies symmetric about the equator for both the EOF1 and EOF2 modes. In the rest of this paper, we will focus on the QBWO in boreal spring.

4. 3D circulation patterns associated with QBWO in boreal spring

A linear-lagged regression is used to reveal the circulation pattern associated with a full cycle of the QBWO. As shown in Fig. 2a, the first two PCs have a significant correlation, with the peak of PC1 leading that of PC2 by 3–4 days. Thus, one may simply choose PC1 as the reference time series and reveal the QBWO evolution with the application of a linear-lagged regression method. Figure 3 illustrates the evolution of the OLR and the low-level (850 hPa) circulation associated with a half cycle of the QBWO from day −7 to day 0. Here, day 0 (day −7) represents the strongest (most suppressed) convective phase of the QBWO over the eastern equatorial IO. Before conducting the lagged regression, both the OLR and 850-hPa wind fields are subjected to a 10–20-day bandpass filter. In the rest of this study, without being mentioned specifically, all discussions are for the QBWO and all regression coefficients are relevant to PC1.

At day −7, a strong positive OLR center, representing the suppressed convection, appears over the eastern equatorial IO. Correspondingly, low-level easterlies are pronounced in situ. A weak negative OLR anomaly occurs in the western IO. In the subsequent days (days −6 to −5), the positive anomaly over the eastern IO tends to separate into two symmetric parts in association with two anticyclonic gyres, whereas the negative anomaly in the west propagates slowly toward the east.

From days −4 to −3, the negative OLR anomaly rapidly shifts into the eastern equatorial IO, while the two positive OLR anomaly centers, along with two anticyclonic gyres, move poleward. An interesting feature of the low-level circulation at this transition phase is the easterly anomaly overlapping with the enhanced convection in the eastern equatorial IO. This structure is contradictory to a conventionally held Gill-type dynamic pattern of the westerly anomaly being in phase with the enhanced convection anomaly. As will be discussed in the next section, the equatorial easterly anomaly (associated with the two off-equatorial anticyclonic gyres) is a cause rather than a result of the local convection anomaly.

Two days later (days −1 and 0) when the equatorial convection is further enhanced and the impact of the off-equatorial gyres is weakened, the low-level westerly anomaly forms and is in phase with the negative OLR center. The spatial patterns of the OLR and 850-hPa winds at day 0 are a mirror image of those at day −7, indicating a completion of a half cycle of the QBWO from days −7 to 0.

One may note from Fig. 3 that the OLR patterns at day 0 (day −4) are similar (opposite) to the first (second) leading EOF patterns shown in Fig. 2a. This again indicates that the two leading EOF modes reflect the different phases of the same dynamic mode.

A key question related to the formation of the QBWO convection in the eastern equatorial IO is whether it is generated locally or attributed to the eastward propagation of the disturbance from the western IO. Note that the negative OLR center is located at 65°E in the western equatorial IO at day −7 and moves to 75°E at day −4, with an averaged phase speed of 3.3° day$^{-1}$. However, at day −3, it suddenly jumps to 90°E, 4–5 times faster than the previous propagation speed. To reveal the cause of this "odd" feature, we examine the longitude–vertical section of the regressed zonal and vertical winds at each phase (Fig. 4).

At day −7, a strong subsidence is pronounced throughout the troposphere over the eastern equatorial IO (80°–100°E), coinciding with the positive OLR anomaly in situ. This subsidence weakens from days −7 to −5 and, meanwhile, a local-ascending branch develops in the planetary boundary layer (PBL), gradually penetrating into the upper levels. To the west of the midtropospheric subsidence, there is an updraft over the western IO, which moves slowly eastward. Its location coincides with that of the negative OLR center (see Fig. 3). At day −4, the western and eastern branches of the updrafts are well connected at low levels, while the downdraft in between weakens and can only be detected at upper levels. At day −3 when the equatorial eastern IO is overwhelmed by large-scale ascending motion, one may still discern two ascending centers, with one being near 70°E and the other near 100°E. The former is
initiated from the western IO, whereas the latter develops locally, as viewed clearly from the vertical–longitude maps. The local ascending motion initially develops at PBL (below 800 hPa), and it grows rapidly and reaches to the middle troposphere at day $-4$. As the eastern convective branch continues growing from days $-3$ to 0, the western one decays gradually.

Therefore, the evolution of the zonal and vertical circulation along the equatorial plane reveals the cause of the sudden “jump” of the OLR anomaly from the western to eastern IO. It provides a physical clue of how the local convection in the eastern IO is generated—it is primarily attributed to the local processes associated with the atmospheric boundary.
layer dynamics. Note that after the low-level westerly is established in response to the local convective heating at day 0, a new downdraft is formed at the boundary layer around 100°E, indicating the subsequent development of a suppressed convective phase over the region.

Figure 5 further illustrates the evolution of the QBWO circulation in the vertical–latitude section along...
100°E. The circulation is approximately symmetric about the equator. As in the vertical–longitude section, the local ascending motion is first found at PBL during the suppressed convective phase over the eastern equatorial IO. It enhances and extends upward. Meanwhile, the downdrafts at both sides of the equator weaken gradually. Therefore, both the vertical–longitude and vertical–latitude profiles reveal that even though there are obvious eastward-propagating signals along the equatorial IO, the formation and phase transition of convection
in the eastern equatorial IO are primarily attributed to local processes. In the next section, we will investigate how the boundary layer ascending motion is generated.

5. Mechanism for the boundary layer triggering off Sumatra

As shown in Figs. 3–5, the ascending motion first develops in the boundary layer at days $-7$ to $-6$ when suppressed convection and descending anomalies appear over the equatorial eastern IO. How does the vertical motion form in the boundary layer? As we know, in the absence of the background mean flow, the free-atmospheric response to a midtropospheric heating/cooling is baroclinic, that is, the upper and lower tropospheric circulations are reversed (Gill 1980). In the presence of the background vertical shear, the free-atmospheric barotropic flow may be generated because of the coupling between the barotropic and baroclinic flows (Wang and Xie 1996; Jiang et al. 2004). The divergence of the free-atmosphere barotropic flow may further induce the boundary layer convergence according to the mass continuation. Thus, it is critical to examine the evolution of the barotropic flow associated with the QBWO, and its phase relationship with the baroclinic flow and PBL divergence.

A dependent variable may be decomposed into a barotropic ($+$) and a baroclinic ($-$) part (Peixoto and Oort 1992) as

$$A(x, y, z, t) = A_\times(x, y, t) + A_\times(x, y, z, t),$$

where $A_\times = \int_{P_t}^{P_t} A dp/(P_b - P_t)$ and $\int_{P_t}^{P_b} A dp = 0$. $P_t$ and $P_b$ are pressures at the top and bottom of the free atmosphere, respectively. Here we set $P_b = 850$ hPa and $P_t = 100$ hPa.

Our diagnosis reveals that at the equator the zonal wind component primarily contributes to the free-atmosphere barotropic divergence field. Figure 6 illustrates the evolution of the free-atmospheric barotropic zonal wind along the equator in a full cycle of the QBWO. The amplitude of the barotropic zonal wind is about 0.5 m s$^{-1}$, being close to that of the baroclinic counterpart. The maximum barotropic easterly occurs at day $-6$ and is located at 90°E. The maximum barotropic easterly appears 1–2 days after the weakest convective phase over the region, implying that the barotropic flow is possibly triggered by the anomalous descending motion associated with the QBWO. Unlike the behavior of the OLR anomaly, the barotropic flow propagates slowly westward along the equator, possibly caused by the Rossby wave emanation from the anticyclonic gyres off the equator (Wang and Xie 1997).

To understand the specific process that generates the barotropic zonal wind, we diagnose the zonal momentum budget using the NCEP–NCAR reanalysis data. All dependent variables are decomposed into a basic state and a perturbation field as

$$A = \bar{A} + A'. $$

![FIG. 6. Free-atmospheric barotropic zonal wind (10$^{-1}$ m s$^{-1}$) along 5°S–5°N regressed onto the PC1 from days $-7$ to 7. The light-filled area denotes easterlies and the dark-filled area denotes westerlies.](image-url)
Substituting Eq. (4) into the zonal momentum equation, considering that the mean state satisfies the equation and integrates vertically from 850 to 100 hPa, one may derive the barotropic zonal wind tendency equation as

\[
\frac{\partial u'}{\partial t} = -U_u - V_u - W_u + F_y + \Phi_x, \tag{5}
\]

where

\[
U_u = \int_{P_1}^{P_2} \left( u \frac{\partial u'}{\partial x} + u' \frac{\partial u}{\partial x} + \omega' \frac{\partial u}{\partial p} \right) dp/(P_b - P_t), \tag{6}
\]

\[
V_u = \int_{P_1}^{P_2} \left( \frac{\partial u'}{\partial y} + u' \frac{\partial u}{\partial y} + \omega' \frac{\partial u}{\partial p} \right) dp/(P_b - P_t), \tag{7}
\]

\[
W_u = \int_{P_1}^{P_2} \left( \frac{\partial u'}{\partial p} + \omega' \frac{\partial u}{\partial p} + \omega \frac{\partial u}{\partial \bar{p}} \right) dp/(P_b - P_t), \tag{8}
\]

\[
F_y = \int_{P_1}^{P_2} f \omega' dp/(P_b - P_t), \tag{9}
\]

and

\[
\Phi_x = \int_{P_1}^{P_2} \frac{\partial \phi'}{\partial x} dp/(P_b - P_t). \tag{10}
\]

For simplicity, we define \( \bar{A} \) as the climatological mean from 1 March to 31 May and \( A' \) as the 10–20-day filtered field.

Figure 7 shows the barotropic zonal wind, and its tendency averaged over the eastern equatorial IO (between 80° and 100°E). Note that the maximum barotropic easterly appears at day \(-6\), while the peak of the barotropic westerly occurs at day 1. To understand the generation mechanism for the barotropic easterly–westerly, we diagnose each term at the right-hand side of Eq. (5) and a composite is made for a 2-day period prior to the peak of the barotropic easterly and westerly.

Figure 8a shows the composite differences between the barotropic easterly and westerly phases. Note that the vertically integrated horizontal advection and pressure gradient force terms are all small. The dominant contribution comes from the vertical advection term. The zonal momentum budget is not exactly in balance, possibly due to finite difference errors, vertical interpolation from a sigma to a pressure coordinate, and/or the neglecting of horizontal and vertical diffusions in the momentum equation. A further discussion of this discrepancy can be found in the next section.

According to Eq. (8), the vertical momentum transport consists of the following three terms: 1) the vertical
advection of the perturbation zonal wind by the background mean vertical motion $W_{u,1}$. 2) the vertical advection of the mean zonal wind by the perturbation vertical motion $W_{u,2}$. and 3) the vertical transportation of the perturbation zonal wind by the perturbation vertical motion $W_{u,3}$, respectively. The physical interpretation of Eq. (8) is that a baroclinic perturbation may interact with either a baroclinic mean flow, or the baroclinic perturbation itself, to lead to the generation of the barotropic motion in the free atmosphere.

The relative contributions of the three vertical advection terms are diagnosed and the result is illustrated in Fig. 8b. It is found that the vertical transportation of the background mean zonal wind by $W_{u,2}$ contributes mostly to the vertical momentum transport, whereas the other two terms play a minor role. As shown in Fig. 4, at day $-7$, strong descending motion occupies the whole troposphere east of $80^\circ$E. The background mean flow in boreal spring over the eastern equatorial IO exhibits an easterly shear feature with low-level westerlies (Zhang et al. 2002; He et al. 2006) and upper-level easterlies (Fig. 9). As a result, the contribution to the tendency of the barotropic zonal wind $-\omega^1 (\partial u^1 / \partial p)$ is negative. Thus, during the strongest suppressed convective phases at days $-7$ and $-6$, the perturbation descending motion advects the mean easterly momentum downward, leading to the generation of the free-atmospheric barotropic easterly. The opposite process is found during the generation of the barotropic westerly.

The result shown previously suggests that the free-atmospheric barotropic motion may be generated through the interaction between the QBWO baroclinic motion and the baroclinic mean flow. How does the generated barotropic flow further impact the boundary layer circulation? Note that the strong barotropic easterly occurs at days $-7$ and $-6$ around $90^\circ$E. To the east of the maximum easterly, the zonal gradient of the barotropic wind is divergent. In addition, the barotropic easterly at the equator may also induce a tendency of the meridional wind divergence (i.e., the $\beta$ divergence) in the free atmosphere as

$$\frac{\partial}{\partial t} \left( \frac{\partial u_+}{\partial y} \right) \approx -\beta u_+. \quad (11)$$

Our calculation indicates that in the eastern equatorial IO the zonal wind component dominates the free-atmospheric divergence field. According to the mass continuity, this free-atmospheric divergence needs be balanced by a boundary layer convergence so that the total column mass divergence vanishes. Figure 10 shows the evolution of the free atmosphere and PBL divergence patterns from days $-7$ to $0$. Note that they in general have opposite signs. The near-surface convergence is approximately in phase with anomalous ascending motion at top of the atmospheric PBL.

The previous argument suggests a possible scenario for the local origin of the QBWO off Sumatra. Prior to day $-6$, there is a pronounced descending anomaly in the equatorial eastern IO in association with the suppressed convective phase of the QBWO. The strong descending motion advects the mean easterly momentum downward, leading to the formation of the free-atmospheric barotropic easterly in situ (Figs. 4 and 6). The barotropic easterly further causes the free-atmospheric divergence and the boundary layer convergence. As a result, a marked updraft is observed first at the top of the boundary layer around $100^\circ$E, and then it penetrates into the upper troposphere (Fig. 4). Thus, it is the local PBL process that leads to the phase change of the QBWO convection in the eastern equatorial IO from a break phase at day $-7$ to an active phase at day $0$. A reversed process is followed as the anomalous convection leads to the formation of the free-atmospheric barotropic westerly.

To demonstrate clearly the phase relationship among the OLR, the barotropic wind, and the boundary layer divergence, we show the time evolution of these fields averaged in the eastern equatorial IO in Fig. 11. Here, the 10–20-day-filtered 925-hPa specific humidity and winds are used to represent the boundary layer moisture and wind fields. Note that the maximum positive OLR anomaly (and anomalous descending motion at 500 hPa) leads the peak of the barotropic easterly by one day
Fig. 10. Free-atmosphere (averaged over 850–100 hPa) and PBL (at 925 hPa) divergence associated with QBWO. The contour interval is $1 \times 10^{-2}$ hPa s$^{-1}$. 
amplitude off the equator, with a typical Rossby wave structure, and propagates westward. The present study shows that the structure and generation mechanism of the boreal spring QBWO are different from those in boreal summer and winter. In boreal spring, QBWO is approximately symmetric about the equator, and it originates from local processes over the eastern equatorial IO, not signals propagating from the western IO. The cause of the structure difference between the boreal spring and summer/winter QBWO is likely attributed to the mean state distribution (the spring mean state is approximately symmetric, whereas the summer–winter mean state is quite asymmetric relative to the equator). A further study is needed to understand the dynamic cause of structure asymmetry and temporal-scale selection.

A relevant question is what differentiates QBWO and MJO in boreal spring? To show their differences, we extract the EOF modes of MJO in boreal spring (MAM) in the same way as we did for the QBWO analysis. Then, we regress the OLR and circulation fields onto PC1 for both the MJO and QBWO modes.

Figure 12 illustrates the regressed OLR and 850-hPa wind fields for both the MJO and QBWO modes in the global tropics domain. It is clearly seen from Fig. 12 that the spatial scale of MJO is much larger than that of QBWO. While dominant convective activity associated with QBWO is confined in the tropical IO and western Pacific warm pool, MJO convection shows a more pronounced global-circumnavigating characteristic. Figure 13 further illustrates the evolution of the 200-hPa velocity potential field associated with the MJO and QBWO modes. While the MJO is dominated by a zonal wavenumber-1 structure, the QBWO has a remarkable wavenumber-2 pattern.

Therefore, the previously stated analysis clearly illustrates the marked differences between QBWO and MJO in spatial structure, zonal wavelength, and propagation speed.

Despite the difference in zonal wavelength and frequency, QBWO and MJO exhibit a similar horizontal Kelvin–Rossby wave couplet pattern. This implies that the fundamental mechanism that causes the eastward propagation for the two modes may be similar.

An interesting feature is the sinusoidal nature of the time variations of QBWO, even though they contain a highly nonlinear moist process, which is characterized by an asymmetric nature associated with upward and downward motions. It is argued that the moist process is nonlinear or asymmetric in association with the total upward and downward motions. However, for the perturbation it will depend on the background mean state under which the perturbation evolves. During boreal spring, the pronounced seasonal mean large-scale convection

6. Discussion

Most of the previous works are focused on QBWO in boreal summer or winter, which has the maximum

(Fig. 11a), whereas the barotropic easterly leads the free-atmospheric divergence by a couple of days (Fig. 11b). The free-atmospheric divergence, on the other hand, is in phase with the boundary layer convergence (Fig. 11b). The near-surface specific humidity increases from days −6 to 1, when the boundary layer flow is convergent (Fig. 11c). The increase of the moisture resulting from the boundary layer convergence may further destabilize the local atmosphere, leading to the upward development of the ascending branch (Fig. 4). The enhanced surface moist static energy eventually leads to a convectively unstable stratification and triggers the onset of convection. Therefore, the boundary layer divergence induced by the free-atmospheric barotropic winds over the eastern equatorial IO is a direct cause of the phase transition of the QBWO.
and upward motion are often observed over the equatorial IO. As a result, anomalous downward (upward) motion may lead to a negative (positive) heating anomaly. This indicates an approximately linear process. However, there is still an asymmetry in the zonal extension between the convective and nonconvective regions—a narrower rising branch is accompanied by a wider subsidence branch. Such a zonal asymmetry, however, was largely reflected by the EOF1 and EOF2 patterns, so that the time series of their principal components remain sinusoidal. It is possible that a bandpass filter may help smooth out some temporally asymmetric features.

The relatively large residual term in the momentum budget analysis is possibly attributed to the neglecting of the variability with time scales longer than 20 days in the background mean flow. The effect of this slowly varying background state was revealed from the composite analysis of Fukutomi and Yasunari (2005) and Yokoi and Satomura (2006). However, in the current work the budget analysis is done at each phase of QBWO, which is derived previously, based on regressed coefficients in boreal spring during the entire 27-yr period. As a result, for each phase there is no specific information about the timing of the background state and only a seasonal mean background state was used in the calculation. Another possible cause is the neglecting of the PBL dissipation.

7. Conclusions

The spatial and temporal characteristics of the QBWO are investigated over the tropical IO using 27-yr daily
OLR and NCEP–NCAR reanalysis data. The QBWO convective activity is closely related to the climatological mean convective activity, with active centers shifting seasonally along with the mean convection centers. There is a permanent QBWO activity center over the equatorial eastern IO. The intensity of QBWO is in general comparable with that of MJO. A double-cell Rossby gyre structure is often observed at low levels in association with the QBWO convective anomaly, whose axis is shifting meridionally with each season. The most equatorially symmetric patterns for both the OLR and wind fields are found in boreal spring over the tropical IO. The detailed 3D structure and evolution characteristics of the symmetric QBWO mode in boreal spring are examined. It is found that the QBWO convection first appears over the western equatorial IO and then moves eastward slowly. After passing the central IO, the disturbance suddenly speeds up and jumps into the eastern equatorial IO. Meanwhile, the preexisting suppressed convective disturbance in the eastern IO shifts poleward, accompanied by low-level anticyclonic Rossby gyres. The evolution of the circulation along the longitude–vertical section reveals that prior to the phase transition of the QBWO from a suppressed to an active convective phase, an ascending motion (and convergence) anomaly emerges first in the atmospheric PBL around 100°E. The increase of the surface moisture (resulting from the PBL convergence) further leads to the convectively unstable stratification and, thus, the upward development of the ascending motion and convection. Therefore, the QBWO convection over the eastern equatorial IO is primarily triggered by local processes in the PBL, rather than resulting from the propagation of convective disturbances from the western IO.

A zonal momentum budget analysis using the NCEP–NCAR reanalysis data is conducted to reveal the cause of the free-atmospheric barotropic flow and associated divergence. The result indicates that the downward transport of the background mean easterly by perturbation vertical motion leads to the generation of the barotropic easterly in the free atmosphere. The divergence of the barotropic easterly to the east of the maximum zonal wind further leads to a convergence and ascending motion at the atmospheric boundary layer. The PBL convergence causes the increase of the surface moisture and, thus, the convective instability in situ. Therefore, through anomalous downward easterly momentum transport, and its associated impact in the atmospheric PBL, a suppressed convective anomaly over the eastern equatorial IO may lead to the onset of a new opposite-phase convective anomaly in situ.

The previously stated observational analysis reveals that the phase transition of the QBWO convection over the eastern equatorial IO is primarily attributed to the local boundary layer process. The eastward propagation of convective disturbances along the equatorial IO plays a minor role in the phase transition.

The study of the QBWO in boreal spring is useful because it may significantly impact the Asian monsoon onset (Zhang et al. 2002; Wen and Zhang 2007). The generation mechanism for QBWO in boreal spring might be different from that in other seasons. How and to what extent does the local triggering mechanism operate in other seasons requires further observational analyses and modeling studies.

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