The Madden–Julian Oscillation Observed during the TOGA COARE IOP: Global View

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ABSTRACT

During the TOGA COARE Intensive Observing Period (November 1992–February 1993), two pronounced Madden–Julian oscillation (MJO) events associated with super cloud clusters and westerly wind bursts were observed. This paper presents a global view of the MJOs including the origin of the super clusters in the Indian Ocean, their migration into the Maritime Continent and the TOGA COARE large-scale soundings array (LSA) in the western equatorial Pacific, and their rapid decay over cold water of the eastern Pacific. The structure and evolution of the MJO are examined with emphasis on the coupling between large-scale motion and convection. Because of differences in propagation speeds, the positions of maximum zonal wind perturbations relative to deep convection undergo systematic changes during the travel of the MJO. However, the centers of deep convection always coincide with those of large-scale ascent. The super cloud cluster accompanies a wide area of warm air in the upper troposphere. Over the warm pool region the perturbation kinetic energy of the motion in the 30–60-day period range is maintained by the conversion of perturbation available potential energy generated by convective heating. Over the central-eastern Pacific, there is strong horizontal convergence of wave energy flux entering the equatorial upper-tropospheric westerly duct from the extratropical latitudes, suggesting interactions of the MJO with midlatitude disturbances.

1. Introduction

The intraseasonal zonal wind oscillation is among the most prominent large-scale motions in the tropical atmosphere. In their pioneering work, Madden and Julian (1971) detected a spectral peak with a period near 40–50 days in time series of zonal wind and surface pressure at Canton Island. Then they showed that the oscillation corresponds to a global-scale zonal circulation cell propagating eastward along the equator (Madden and Julian 1972). From the analysis of water vapor mixing ratios and temperature, they inferred that the circulation cell is coupled with enhanced large-scale convection in the sector from the Indian Ocean to the western Pacific Ocean.

Zangvil (1975) and Yasunari (1979, 1980) showed evidence of large-scale eastward moving disturbances in the satellite-measured cloud field over the Indian Ocean. Zangvil and Yanai (1981) noted high coherence between the satellite-measured brightness and long-period eastward moving disturbance of the zonal wind at 200 hPa. The deduced structure of the disturbance was consistent with that of a global oscillation envisioned by Madden and Julian (1972). Since then, the Madden–Julian oscillation (hereafter MJO) has been a focus of intense research. Numerous papers have been published to describe the spatial and temporal characteristics of the MJO [see Madden and Julian (1994) for a review].

Several theories have been proposed to explain the origin and characteristics of the MJO. For example, the response to thermal forcing (e.g., Yamagata and Hayashi 1984; Salby and Garcia 1987; Salby et al. 1994), the instability of moist Kelvin waves through wave-CISK (conditional instability of the second kind) (e.g., Lau and Peng 1987; Wang 1988; Lim et al. 1990), or through evaporation–wind feedback (e.g., Emanuel 1987; Yano and Emanuel 1991; Neelin and Yu 1994) [see Hayashi and Golder (1997) for an extensive review]. However, these theories have not been fully tested or verified against observations. Many attempts to simulate the MJO by numerical models have met with difficulties in producing the MJO or obtaining realistic propagation speed of the MJO-like oscillation appearing in simulation (e.g., Hayashi and Sumi 1986; Swinbank et al. 1988; Slingo et al. 1996). Even the results from more recent model simulations with various degrees of sophistication (e.g., Hayashi and Golder 1997; Sperber et
A key question in the dynamics of the MJO is the manner in which cumulus convection and the wind oscillation interact with each other. The eastward moving convection associated with the MJO appears to originate in the Indian Ocean and decays as it reaches the colder sea surface temperature (SST) region near the date line. Nevertheless, the upper-tropospheric portion of the zonal wind oscillation appears to continue its eastward propagation. Because of these features, the circulation anomalies associated with the MJO have been classified into two regimes: a convective regime across the Indian and western Pacific Oceans, and a dry regime elsewhere (Knutson et al. 1986; Gutzler and Madden 1989; Salby and Hendon 1994). The large-scale circulation anomalies in the upper troposphere propagate eastward at a speed of ~5 m s\(^{-1}\) in the convective regime and at a speed ranging from 10 to 15 m s\(^{-1}\) in the dry regime (Knutson et al. 1986; Hendon and Salby 1994).

Recently, Milliff and Madden (1996) and Milliff et al. (1998) detected eastward propagating signals in station surface pressure and the surface zonal wind in the European Remote Sensing Satellite (ERS-1) data with speeds of 30–40 m s\(^{-1}\) in the equatorial eastern Pacific. They conceptualized this as a far-field dispersion product of strong convection associated with the MJO in the Indian Ocean and the western Pacific. Bantzer and Wallace (1996) found fast eastward propagating signals in upper-tropospheric temperature and zonal wind, emanating from the region of enhanced precipitation associated with the 40–50 day oscillation.

Furthermore, in the convective regime, satellite observations revealed the existence of convective organization of various space scales and timescales (e.g., Nakazawa 1988; Lau et al. 1991; Sui and Lau 1992). Nakazawa (1988) found the super cloud cluster structure in the satellite-observed cloud patterns in which a large-scale eastward propagating envelope of organized convection is composed of westward moving individual cloud clusters. Westward propagating disturbances with a periodicity of ~2 days have been noted by Takayabu (1994), Chen et al. (1996), and Haertel and Johnson (1998).

The Intensive Observing Period (IOP) of the Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean–Atmosphere Response Experiment (COARE) took place from 1 November 1992 through 28 February 1993. One of the stated goals of TOGA COARE was to determine the structure of the synoptic and mesoscale components of the large-scale slowly varying atmospheric circulation over the warm pool and, in particular, to determine the morphology of convective stages of the 30–50 days oscillation and its subcomponents (Webster and Lukas 1992).

In this paper, we present a global view of the two prominent MJO events observed during the TOGA COARE IOP. The first objective of this case study is to describe the structure and evolution of the MJOs with emphasis on the coupling between large-scale wind perturbations and deep cumulus convection. The second objective is to examine the energy transformation processes associated with the intraseasonal (30–60 day) variability in the equatorial belt containing the MJO.

In section 2 the datasets and analysis procedures are briefly described. Section 3 presents the mean global distributions of the horizontal and vertical large-scale circulation during TOGA COARE IOP as backgrounds. Section 4 discusses the propagation of envelopes of super cloud clusters as seen in the outgoing longwave radiation (OLR) field and associated zonal wind and temperature perturbations. This section also examines the structure of the MJO and its time evolution with emphasis on coupling between convection and the large-scale motion. Section 5 discusses the maintenance mechanisms of kinetic energy associated with the 30–60-day variability in the equatorial belt. Section 6 presents a summary and discussion.

2. Data and analysis methods

In this study, the European Centre for Medium-Range Weather Forecasts reanalysis (ERA) and the “interpolated” OLR datasets from November 1992 to February 93 are used. The ERA was generated by a frozen global data assimilation system along with a database, which is as complete as possible for the period 1979–93 (Gibson et al. 1997). The interpolated OLR dataset is a set in which the missing data are filled by using temporal and spatial interpolation (Liebmann and Smith 1996). Because the interpolated OLR data are available only once a day, the original ERA made four times a day (0000, 0600, 1200, and 1800 UTC) are averaged to generate the set of daily mean data. Both datasets are on a 2.5° × 2.5° grid.

Main diagnostic tools used in this case study are simple horizontal maps and longitude–time and longitude–height sections. In addition, cross-spectral analyses are applied to the OLR, wind, vertical motion, temperature, and heating rate to examine the manner of coupling between convection and the large-scale motion, and to
estimate the terms involved in the energy transformation processes associated with the intraseasonal (30–60 day) variability.

3. Global features of mean circulation and convective activity during IOP

a. Mean circulations

In Fig. 1, the IOP-mean streamline maps at 150 and 850 hPa are illustrated. At 150 hPa (Fig. 1a) a pair of huge anticyclonic circulations with an axis of diffuence along the equator is prominent over the longitudes from the Indian Ocean to the date line. A pronounced outflow center is seen near the date line where convection is most intense (Gutzler et al. 1994). In addition, anticyclones are located over southern Africa and South America where convection is active. In Fig. 1a we also note that easterly winds prevail over the equatorial belt from 180° to Africa, while westerly winds are seen over the equatorial eastern Pacific and Atlantic. The mean wind in the Tropics at 850 hPa (Fig. 1b) generally flows in the opposite direction to the winds at 150 hPa. There is a large-scale cross-equatorial flow over the Indonesian Archipelago connecting the Asian winter monsoon to the Australian summer monsoon characterized by the equatorial westerlies (e.g., McBride et al. 1995).
Figure 2a shows the longitude–height section of the IOP mean zonal wind on the equator. The mean zonal flow from the eastern Indian Ocean to near the date line is westerly in the lower troposphere and easterly in the upper troposphere. This vertical distribution of the mean zonal flow is reversed over the longitudes covering the central-eastern Pacific and Atlantic Oceans where deep convection is suppressed. We note two pronounced centers of the upper-tropospheric westerlies centered at 150 hPa over the central-eastern Pacific and Atlantic. The association of the upper-level easterlies with active convection, and that of the upper-level westerlies with suppressed convection have been well recognized (e.g., Liebmann 1987). The IOP mean vertical circulation in the longitude–height section along the equator is shown in Fig. 2b. The vertical velocity, \( w \), is obtained by converting the vertical \( \rho \) velocity, \( \omega \). Mean upward motions occupy a wide region from the Indian Ocean (∼80°E)
to the central Pacific west of 150°W, and over South America and Africa. The areas of mean upward motions are quite consistent with those of active convection reflected in the OLR field (not shown). The maximum upward velocities are located over the eastern Indian Ocean (~90°E) and just west of the date line.

b. Comparisons with climatology

The COARE IOP occurred during a lingering period of warm ENSO episode (e.g., Gutzler et al. 1994). In a separate paper Chen and Yanai (2000) have compared the SST and the intensity of the 30–60-day oscillations during the COARE IOP with a 15-yr (1979–93) climatology. The SST over the regions of the Intertropical convergence zone (ITCZ) of the eastern Pacific and the South Pacific convergence zone (SPCZ) during the IOP was slightly higher than the 15-yr mean SST of these regions by about 0.2°C, and SST along the coast of California was warmer by more than 1°C. However the COARE Outer Sounding Array (OSA) was covered by small negative SST anomalies that were responsible for relatively weak convection in this region. The comparison of the intensity of 30–60-day oscillations show that during the COARE IOP the intraseasonal variation was intensified near the date line.

4. Propagation and time evolution of super cloud clusters and associated perturbations

a. The life cycle of super cloud clusters

Figure 3 shows the longitude–time section of the OLR averaged between 5°S and 5°N. As noted previously by Velden and Young (1994) and Nakazawa (1995), two distinct envelopes of super cloud clusters (hereafter called “super clusters” for brevity), denoted by letters A and B, propagating eastward from the Indian Ocean to the vicinity of the date line are recognized in the IOP. The super cloud cluster A started to propagate eastward from the equatorial Indian Ocean near 75°E in late November 1992. This super cluster crossed the Indonesian Maritime Continent and reached the large-scale soundings array (LSA) in the western Pacific, then disappeared near 170°W around 10 January 1993. The second cluster B began to migrate from the Indian Ocean around 5 January and reached the central Pacific in early February.

Note that super cloud clusters existed only over the warm water of the Indian and western Pacific Oceans where the mean vertical motion is upward (Fig. 2b). The eastward propagating speed of the super cluster was approximately 5 m s⁻¹ although local variations in speed were evident. Standing convection is also observed between 90° and 120°E before and after the two events of propagating super cloud clusters. In addition, two regions of stationary convection over equatorial Africa and America are seen in Fig. 3.

b. Propagation of zonal wind and temperature perturbations

The tropical large-scale wind field exhibited multiscale temporal and spatial variabilities during the IOP. Figures 4a and 4b illustrate the longitude–time sections of zonal wind perturbations at 850 and 150 hPa, respectively, along the equatorial belt. In this subsection, the perturbation of zonal wind velocity is calculated by subtracting the zonal and time means from the original data to represent transient wave disturbances, that is,

\[ u^* = u - \bar{u} - u^* - [u'] \]

(1)

where \( \bar{u} \) stands for the zonal mean, \( (\cdot) \) the deviation from the zonal mean, \( \bar{t} \) the time mean, and \( (\cdot)' \) the deviation from the time mean (e.g., Peixoto and Oort 1992, 61–62).

At 850 hPa eastward moving bands of westerly and easterly (shaded) perturbations alternating with intraseasonal timescales are evident over the warm water of the Indian and the western Pacific Oceans (Fig. 4a). Two bands of anomalous westerlies traveled over the warm water pool in two periods (early December to early January, and mid-January to mid-February). These bands brought two westerly wind bursts (WWBs) near the date line. Both bands did not extend much beyond the date line. From Figs. 3 and 4a, we recognize that the centers of active convection lead the WWBs in the western Pacific, as reported by many investigators (e.g., Sui and Lau 1992; Kiladis et al. 1994; Lin and Johnson 1996a). However, as the convection slows down significantly around the date line (e.g., around 25 December and 5 February), the WWBs occur concurrently with the centers of active convection. Outside the warm pool region, westward moving perturbations with shorter timescales are more evident especially in November–December. The westward moving waves over the eastern Pacific have been identified as equatorially trapped Rossby waves (Kiladis and Wheeler 1995; Meehl et al. 1996; Pires et al. 1997).

At 150 hPa two eastward moving bands of westerly perturbations and two eastward moving bands of easterly perturbations associated with the MJO are clearly recognized (Fig. 4b). The overall amplitude of the zonal wind perturbation at 150 hPa, \( u^*(150) \), is about twice as large as that of at 850 hPa, \( u^*(850) \). Furthermore, \( u^*(150) \) continues to propagate beyond the date line with increased amplitude over the eastern Pacific where deep convection is suppressed. An interruption of its eastward propagation is noted in longitudes centered around the Andes Mountains (~80°W). In addition, there is a suggestion of occasional westward propagation (e.g., 5–25 January). The eastward propagation resumes over the Atlantic Ocean with a very fast phase speed (~20 m s⁻¹). As a result, around 7 January, the super cluster B was intercepted by a fast-moving band of easterly perturbation at 150 hPa, which was, in fact, a remnant of the previous MJO.
Fig. 3. Longitude-time section of OLR (W m$^{-2}$) averaged between 5°S and 5°N (contour interval: 15 W m$^{-2}$). Areas with OLR values less than 215 W m$^{-2}$ are shaded. The origin of two envelopes of super cloud clusters A and B are labeled, respectively.
Fig. 4. Longitude–time sections of zonal wind perturbation $u^*$ (m s$^{-1}$) [see Eq. (1)] averaged between 5$^\circ$S and 5$^\circ$N: (a) 850 and (b) 150 hPa. Areas of negative values (easterly perturbations) are shaded. The 170 W m$^{-2}$ contours of OLR are also shown.
FIG. 4. (Continued)
The 400-hPa temperature anomaly, $T^u(400)$, as seen in Fig. 5, propagates with nearly the same speed as the super cloud clusters in the warm water region. Outside the warm pool, however, it moves at the same speed as $u^u(150)$ with large amplitude appearing in the convection-free eastern Pacific around 1 January and 5 February. Over the warm pool, it is seen that the center of deep convection is embedded in the warm anomaly. The center of warm anomaly, however, tends to slightly lead the deep convection. Over the Maritime Continent (110°–150°E) in particular, the warm anomaly marks a distinct signal in the front edge of the eastward propagating convection and persists even after the convection is disrupted. Examples of these are found near 130°E around 5 December and 15 January.

Examining Figs. 3, 4, and 5, we recognize that “preceding disturbances” in zonal wind and temperature perturbations showed up earlier than the super cloud clusters appeared in the OLR field. Around 16 November, for example, amplification of 400-hPa eastward moving warm anomalies can be seen around 40°E, where deep convection is not usually favored. Westerly perturbation at 850 hPa and easterly perturbation at 150 hPa amplify at the same location about 10 days later and move on to the east. The signals in zonal wind and temperature perturbations of the second MJO event can even be traced back to 40°W in early January.

c. Coupling of convection with the large-scale motion

Over the warm pool, the bands of easterly perturbations at 150 hPa (Fig. 4b) are located above the bands of westerly perturbations at 850 hPa (Fig. 4a). When compared with Fig. 3, we recognize that the zonal wind perturbations at both levels propagate faster than the super cloud clusters seen in the OLR field. Because of their different propagation speeds, the position of the super cloud clusters relative to the upper-level and low-level zonal wind perturbations undergoes a systematic change over the warm pool. The super cloud clusters, during their growth to maturity over the western Pacific, are located at the front edges of the low-level westerly perturbation and the upper-level easterly perturbation. When the super clusters approach the date line, however, the low-level WWBs occur simultaneously with the active convection (Fig. 4a). What is the relationship between convection and the horizontal divergence/convergence associated with the MJO at this stage?

In Fig. 6, the coherence square and phase difference of OLR and the vertical $p$-velocity at 400 hPa, $u^u(400)$, at the 40-day period are plotted along the equator for the longitudes covering the warm pool where the super cloud clusters were observed (Fig. 3). In this diagram, the phase difference is defined as positive when the oscillation in $u^u(400)$ leads that of OLR. As we see in the figure, the coherence square is generally high over the warm pool, exceeding the 95% significance level (0.63). Its values drop remarkably, however, over the Maritime Continent (110°–130°E), suggesting a disruption of coupling between convection and the large-scale vertical motion in this period range. The phase difference over the warm pool is nearly zero when the coherence square is large, showing that the centers of deep convection always coincide with those of large-scale ascent regardless of the apparent change of the zonal wind perturbation–OLR relationship seen in Figs. 4a and 4b. This implies that the contribution from $\omega u^u/\omega x$ to the horizontal divergence. Near 70°E in the Indian Ocean, however, there is a suggestion that the large-scale ascent precedes the occurrence of organized convection as seen by the positive phase difference.

d. Three-dimensional structure of the MJO

Figures 7 and 8 illustrate mean maps showing the streamlines of wind at 150 hPa with OLR (top figures) and the streamlines at 850 hPa with the temperature at 400 hPa (bottom figures) for two 5-day periods. The first corresponds to the period when the super cloud cluster A identified in Fig. 3 is located over the Indian Ocean (Fig. 7), and the second to the period when the super cluster A is over the western Pacific (Fig. 8).

The 150-hPa streamlines (Figs. 7a and 8a) show that the easterly flow over the equator associated with the super cloud cluster is accompanied by a pair of anticyclonic circulations ($twin anticyclones$) on both sides of the equator. The super cluster continues to expand until it covers a large area (10°S–20°N, 155°E–170°W) near the end of December. In early January the super cluster rapidly diminishes over the eastern Pacific (not shown).

At 850 hPa (Figs. 7b and 8b) the zonal wind on the equator in the vicinity of the super cloud cluster is typically westerly and it is accompanied by a pair of cyclonic circulations ($twin cyclones$) on both sides of the equator. The low-level twin cyclones are, however, not as conspicuous as the upper-level twin anticyclones. During the slow movement of the super cluster over the Indian Ocean, the upper- and lower-tropospheric flow patterns resemble those obtained as responses to thermal forcing on the equator (Matsumo 1966; Webster 1972; Gill 1980; Wang and Rui 1990; Hendon and Salby 1996). However, the low-level circulation near the super cluster after the MJO leaves the Indian Ocean is more complicated, possibly because of the topography and thermal influence of the Maritime Continent and Australia.

e. OLR–temperature relationship

From a composite analysis of OLR, wind and satellite-measured temperature, Hendon and Salby (1994) showed that, when the MJO is amplifying, active convection is correlated to positive temperature perturbation. They also noted a positive correlation between...
Fig. 5. Longitude–time section of temperature perturbation $T^*$ (K) at 400 hPa. Areas of negative values are shaded. The 170 W m$^{-2}$ contours of OLR are also shown.
upper-tropospheric divergence and temperature. From these they suggested that production of eddy available potential energy and its conversion into eddy kinetic energy occurred in the developing phase of the MJO. Comparing the OLR (Figs. 7a, 8a) and the temperature at 400 hPa (Figs. 7b, 8b), we recognize that the convective region of the MJO is accompanied by a wide region of warm air (see also Fig. 5). Also, compared with an earlier stage (Fig. 7), the warm area associated with the convection expanded as the whole system approached the date line (Fig. 8). Because the low OLR values in the Tropics are very well correlated to the atmospheric heating due to cumulus convection (e.g., Li and Yanai 1996), the observed low OLR–high temperature association implies positive correlation between heating and temperature, and, therefore the production of eddy available potential energy (section 5).

5. Maintenance of kinetic energy associated with the 30–60-day variance

a. Energy equations

To discuss the energy transformation processes associated with the MJO, which propagates in nonuniform flow, it is convenient to consider the equations governing the perturbation kinetic energy (PKE), defined by

\[ \bar{k} = \frac{u'^2 + v'^2}{2} \]  \hspace{1cm} (2)

and a quantity defined by

\[ \bar{e} = \frac{a^2/2S}{(2\pi \alpha \ln \rho)} \]  \hspace{1cm} (3)

where \( a \) is the specific volume, \( S = \frac{-\pi \alpha \ln \rho}{2\pi} \) is the static stability factor, and \( \theta \) the potential temperature. Here \( \bar{e} \) may be termed the perturbation available potential energy (PAPE).

We can show that with the quasi-static approximation,

\[
\frac{\partial \bar{k}}{\partial t} = -\bar{v}' \cdot \bar{\nabla} \cdot \bar{v} - \bar{v}' \cdot \bar{\omega}' \cdot \frac{\partial \bar{\omega}}{\partial \rho} - \alpha \omega' - \bar{\nabla} \cdot \bar{F}_h - \frac{\partial F'_{\rho}}{\partial \rho} + \bar{v}' \cdot \bar{F}'_{\rho}, \]

\hspace{1cm} (4)

where

\[
-\bar{v}' \cdot \bar{\nabla} \cdot \bar{v} = -\left[ \frac{u'' u''}{\partial x} + u'' u' \left( \frac{\partial \bar{\omega}}{\partial y} + \frac{\partial \bar{\omega}}{\partial x} \right) + u' v' \frac{\partial \bar{\omega}}{\partial y} \right],
\]

\hspace{1cm} (5)

and

\[ F_h = \frac{\partial v' v'}{\partial x} + \bar{k} \bar{v} + \frac{1}{2} \bar{v}'^2 \bar{v}' \]  \hspace{1cm} and

\[ F_{\rho} = \frac{\partial \bar{\omega}'}{\partial x} + \bar{k} \bar{\omega} + \frac{1}{2} \bar{v}'^2 \bar{\omega}'. \]  \hspace{1cm} (6)
In (4)–(6), \( v \) is the horizontal velocity, \( \omega = \frac{dp}{dt} \) (the vertical \( p \) velocity), \( p \) the pressure, \( \nabla \) the isobaric gradient operator, \( \phi \) the geopotential, \( f \) the frictional force per unit mass. Here \( F_s \) and \( F_v \) are, respectively, the horizontal and vertical components of the wave energy flux consisting of pressure work and the transports of PKE by the mean flow and perturbations. Equation (4) is analogous to the turbulence kinetic energy equation obtained with the Boussinesq approximation [e.g., Stull 1988, Eq. (5.1a)], although the timescales and space scales of the motion under consideration are vastly different from those of microscale turbulence.

For the quasi-static motion, we can also derive an approximate equation for the time change of PAPE:

\[
\frac{\partial \sigma}{\partial t} = -\frac{1}{S} \alpha^2 \nabla \cdot \nabla \sigma + \frac{1}{2S} \nabla \cdot \sigma - \nabla \cdot E_h - \frac{\partial E_p}{\partial p} + \frac{R}{c_s^2 S_p} \nabla \sigma, \tag{7}
\]

with

\[
E_h = \bar{\sigma} \nabla \cdot \left( \frac{\alpha^2}{2S} \nabla \sigma \right) \quad \text{and} \quad E_p = \bar{\sigma} \nabla \sigma + \left( \frac{\alpha^2}{2S} \right) \nabla \omega, \tag{8}
\]

where \( Q_1 \) is the diabatic heating rate. Equation (7) is similar to the equation for the potential temperature variance due to microscale turbulence [e.g., Stull 1988, Eq. (4.3.3)]. Slightly different forms of (4) and (7) were used previously by Lau and Lau (1992).

The right-hand side terms of (4) represent, respectively, the production of PKE through the work done by horizontal and vertical shear of the mean flow (the shear generation terms), the conversion from PAPE through the \((\alpha, -\omega)\) correlation, the horizontal and vertical convergence of wave energy flux \((F_s, F_v)\), and the work done by frictional force. The right-hand side terms of (7) represent, respectively, the conversion of mean available potential energy through horizontal heat flux due to perturbation, the conversion from PAPE into PKE through the \((\alpha, -\omega)\) correlation, the horizontal and vertical convergence of fluxes of PKE and the production of PAPE through the \((\alpha, Q_1)\) correlation. The second term on the right-hand side of (4) involving the vertical flux of momentum is very small for large-scale motions. The term involving the horizontal flux of heat due to perturbation in (7) is also small in the Tropics (e.g., Zangvil and Yanai 1980).

Therefore, the maintenance of PKE above the boundary layer can be discussed by evaluating the shear generation term that involves only the horizontal flux of momentum (the barotropic conversion term), the conversion from PAPE through the \((\alpha, -\omega)\) correlation, and the horizontal and vertical convergence of wave energy flux in (4). Likewise, the maintenance of PAPE can be discussed in terms of its production due to diabatic heating, \( Q_1 \), and its conversion into PKE and the horizontal and vertical convergence of the PAPE fluxes in (7).

Terms on the right-hand sides of (4) and (7), with exception of the frictional work, can be calculated by using the ERA data. Here \( Q_1 \) is evaluated as the residual of the large-scale heat budget (Yanai et al. 1973). In this work the PKE and all covariance terms of (4) and the \((\alpha, Q_1)\) covariance in (7) were obtained integrating the power spectra and cospectra over the period range of 30–60 days. We found that the sum of the right-hand side terms (except the friction term) of (4) is very small, showing good accuracy of estimation of each term.

b. Distribution of perturbation kinetic energy

Figure 9 illustrates the longitude-height distribution of the IOP mean PKE contained in the 30–60-day period range along the equator. In this figure we notice several major areas of large PKE occupying the equatorial upper troposphere. Areas of large PKE centered at 150 hPa are seen above the Indian Ocean–western Pacific warm pool (70°E–180°). Two deeper layers of large PKE centered at 200 hPa are located over the eastern Pacific and Atlantic Oceans. A relatively small area of PKE exists near 45°E over Africa. In addition, there is a concentration of PKE in the lower troposphere near 170°E, which is a reflection of the westerly wind bursts (section 4). The two PKE maxima over the eastern Pacific and Atlantic Oceans are collocated with the equatorial westerlies in the upper troposphere (Fig. 2a). Many authors have shown that the maximum PKE at 200 hPa in the Tropics is collocated with a zone of equatorial westerlies where convective activity is minimum (Murakami and Unninayar 1977; Arkin and Webster 1985; Liebmann 1987; Webster and Chang 1988). We note, however, that the PKE maxima over the warm pool region are located above the 200-hPa level and were not discussed before.

At least three different hypotheses have been proposed to explain the collocation of PKE maxima with the equatorial westerlies in the upper troposphere. The first considers PKE maxima a result of the accumulation of equatorward wave energy flux in the westerly duct (Webster and Holton 1982; Magana and Yanai 1991; Zhang and Webster 1992). The second hypothesis attempts to explain the formation of PKE maxima as a result of the accumulation of wave energy flux emanation from the tropical source region through nonuniform zonal flow (Webster and Chang 1988; Chang and Webster 1990, 1995). Thirdly, Wang and Xie (1996) have advanced a theory in which the westerly vertical shear provides a favorable condition for the trapping of equatorial waves in the upper troposphere.

c. Production and conversion of perturbation available potential energy

Before the discovery of the MJO, pioneering attempts had been made to obtain a picture of the energy cycle of tropical wave disturbances using sparse upper-station data in the tropical Pacific. Combining the mass and
Fig. 7. The 5-day mean maps for 25–29 Nov 1992: (a) 150-hPa streamlines and OLR (W m⁻²), and (b) 850-hPa streamlines and 400-hPa temperature (K).
Fig. 8. As in Fig. 7 except for 11–15 Dec 1992.
heat budgets and the cospectral analysis, Nitta (1970, 1972) showed that in the Marshall Islands area, the disturbances gain PKE through the conversion of PAPE through positive ($\alpha$, $-\omega$) correlation and that PAPE is generated by cumulus heating through positive ($\alpha$, $Q_1$) correlation. He also found that the correlation terms are large both at the synoptic timescale (4–5 days) and longer timescales (20–30 days). Nitta’s results for the western Pacific were confirmed by Wallace (1971) and Kung and Merritt (1974), and more recently by Lau and Lau (1992). Stevens et al. (1997) have shown that during the TOGA COARE IOP temperature and vertical velocity covariance is positive in the upper troposphere over the IFA. For the intraseasonal (30–50 day) period range, Murakami and Nakazawa (1985) and Krishnamurti et al. (1985) calculated the ($\alpha$, $-\omega$) covariance term using the First GARP (Global Atmosphere Research Program) Global Experiment level IIIb data.

Figure 10 shows the longitude-height cross section of the partial covariance of specific volume ($\alpha$) and vertical motion ($-\omega$) in the 30–60-day period range. The largest energy conversion from PAPE to PKE occurs in a deep tropospheric layer with maximum values at 300–400 hPa over the Indian Ocean–western Pacific warm pool and near the date line, reflecting intense cumulus convective activity inferred from Figs. 2b and 3. Similarly, Fig. 11 shows that the production of PAPE in the same period range occurs over the warm pool due to convective heating through positive ($\alpha$, $Q_1$) correlation. Interestingly, for both terms, there are gaps created by relatively small ($-\omega$, $\alpha$) and ($\alpha$, $Q_1$) covariances over the Maritime Continent (centered near 120°E) where coupling between convection and large-scale vertical motion is weak (Fig. 6). Figures 10 and 11 also show that the production and conversion of PAPE through these correlations are small everywhere except over the warm pool region.

d. Barotropic conversion

A notable feature of the equatorial PKE distribution (Fig. 9) is the areas of PKE maxima in the upper troposphere over the eastern Pacific and Atlantic Oceans where deep convection is suppressed. In the longitude-time section of the zonal wind perturbation $u^{*}$ at 150 hPa (Fig. 4b), large amplitude of $u^{*}$ was observed over the tropical eastern Pacific. We find that the barotropic conversion term (5), which expresses the generation of PKE due to horizontal shear, has large positive values near 160°W and 80°W in the eastern Pacific and at 20°W–0° in the Atlantic (Fig. 12). Previously, Murakami and Nakazawa (1985) calculated the barotropic conversion term (5) at 200 hPa using a bandpass filtered data (centered at 45 day) for the 1979 northern summer and found large positive values over the tropical eastern Pacific (80°–135°W).

e. Wave energy flux and its convergence

To find clues for connecting the production of PKE through the baroclinic and barotropic conversion (Figs. 10 and 12) with the spatial distribution of the observed
Fig. 10. Longitude–height section of the $(\alpha, -\omega)$ covariance in the 30–60-day period range at the equator (J kg$^{-1}$ day$^{-1}$) [see Eq. (4)].

Fig. 11. Longitude–height section of the $(\alpha, Q_1)$ covariance in the 30–60-day period range at the equator (J kg$^{-1}$ day$^{-1}$) [see Eq. (7)].

PKE in the 30–60-day period range (Fig. 9), we construct a longitude–height cross section of the wave energy flux $(F_x, F_z)$ at the equator (Fig. 13) and a horizontal map showing the wave energy flux $(F_x, F_x)$ at 150 hPa (Fig. 14), both for this period range. Here $F_z$ is hydrostatically translated from $F_p$. In addition, we show the longitude–height cross sections of the horizontal and vertical convergence of wave energy flux in Figs. 15a and 15b, respectively.

Figure 13 shows the wave energy flux $(F_x, F_z)$ on the
Fig. 12. Longitude–height section of the barotropic conversion term (5) in the 30–60-day period range at the equator (J kg⁻¹ day⁻¹).

Fig. 13. Longitude–height section of the wave energy flux (F_x, F_z) in the 30–60-day period range at the equator (J m⁻¹ s⁻¹ kg⁻¹) [see Eq. (6)]. Areas of positive (α, −ω) covariance (Fig. 10) are shown by shades.
equatorial plane, where the areas of positive \((\alpha, -\omega)\) covariance are superimposed by shades. In this figure we clearly recognize the wave energy flux “radiated” upward and downward from the source region \((\sim 250-400 \text{ hPa})\) over the warm pool from the eastern Indian Ocean to near the date line where the \((\alpha, -\omega)\) covariance is large positive. Vertical convergence of the upward wave energy flux emanated from the source region appears to be the primary cause of the large PKE in the upper troposphere of the warm pool region near the date line (Figs. 15b). We notice also that the horizontal wave energy flux at 150 hPa (Fig. 14) is eastward from central America to the western Pacific and westward in the central Pacific, contributing to horizontal convergence of the wave energy flow in the upper troposphere near \(170^\circ\text{E}\) (Fig. 15a).

The two pronounced PKE maxima associated with the equatorial westerlies over the eastern Pacific and Atlantic Oceans do not appear to be directly related to the production of PAPE due to convection and its conversion over the warm pool. Instead, the association of these PKE maxima with the barotropic conversion term (Fig. 12) was noted. In addition, examinations of the horizontal as well as vertical wave energy flows reveal additional factors that may be responsible for the PKE maxima in these regions. Figure 14 clearly shows the horizontal wave energy flows entering the equatorial westerly belt over the eastern Pacific from the subtropical latitudes of both hemispheres. The significance of equatorward wave energy flux entering the westerly duct region is seen by the large values of horizontal convergence centered at 200 hPa over the central Pacific (Fig. 15a).

6. Summary and discussion

In this paper we presented a detailed study of the two Madden–Julian oscillation (MJO) events observed during the TOGA COARE IOP. Naturally our conclusions are limited by the small number of cases. Nevertheless, detailed examinations of the data yielded new findings and suggestions for further investigations. The overall representativeness of the intraseasonal variability during the IOP in comparison with a 15-yr (1979–93) climatology is discussed in another paper (Chen and Yanai 2000).

a. Summary

The major findings from this study can be summarized as follows:

1) Structure and evolution

- The super cloud cluster associated with the two MJO events existed only over the warm water pool from \(70^\circ\text{E}\) (the central Indian Ocean) to \(170^\circ\text{W}\) (the central Pacific). They migrated from the central Indian Ocean slowly eastward and were weakened over the maritime continent. After regaining intensity over the western Pacific, they reached maximum intensity near the date line.
On the equator, the eastward moving bands of zonal wind perturbation, $u^\prime$, at 850 hPa were seen only over the warm pool. In the “dry” regime, westward moving Rossby waves were observed at this level.

The $u^\prime$ at 150 hPa and temperature perturbation, $T^\prime$, at 400 hPa propagated nearly “globally,” although an interruption of propagation near the Andes Mountains was noted. Their maximum amplitudes were observed over the eastern Pacific where deep convection was absent.

Over the warm pool, the $T^\prime$ at 400 hPa propagated with about the same speed as that of the super cloud.
clusters (∼5 m s⁻¹), while \( u^* \) at both 150 and 850 hPa propagated faster. In the “dry” regime, \( u^* (150) \) and \( T^* (400) \) had the same fast eastward phase speed as (∼20 m s⁻¹). As a result, the super cluster \( B \) was “intercepted” by a fast-moving band of easterly perturbation at 150 hPa, which was, in fact, a remnant of the previous MJO.

- The signal of the MJO in zonal wind and upper-tropospheric temperature perturbations were found over the equatorial western Indian Ocean, where the convection was suppressed. The super cloud clusters, then, suddenly appeared over the central Indian Ocean. When the envelope of the super cloud clusters was developing to its maturity, it led the eastward moving easterly perturbation at 150 hPa and the westerly perturbation at 850 hPa. On the other hand, active convection was in phase with \( u^* \) at 850 hPa and out of phase with \( u^* \) at 150 hPa near the date line.
- Despite this changing phase relation between the zonal wind perturbation and active convection during the evolution of the MJO, the center of active convection is always located at the center of large-scale ascent. This implies that judging the kinematic properties of the MJO from the zonal wind component alone is misleading and the term \( \partial u/\partial y \) is as important as \( \partial u/\partial x \) to the horizontal divergence. A similar conclusion has been reported for the structure of the MJO obtained in a recent numerical simulation (Matthews et al. 1999). The center of active convection is accompanied by a widespread area of warm air in the upper troposphere.
- The Indonesian Maritime Continent disrupted the convection–motion coupling. This was also seen in the distributions of energy production and conversion terms. The possible cause may be the suppressing effect of diurnal variation of cumulus convection as discussed by Zhao and Weare (1994).

2) Energy Transformation Processes

- The energy cycle deduced from the \((\alpha, Q_1)\) and \((\alpha, -\omega)\) correlations shows that the interaction between convection and large-scale circulation, via production and conversion of PAPE, plays a major role in the maintenance and growth of the MJO over the warm pool. In short, the MJO is maintained by cumulus convection, which is, in turn, sustained by high SST of the warm pool.
- Over the Indian Ocean–western Pacific warm pool, the wave energy flux is clearly radiating upward and downward from the convective source region where \(-\alpha \omega \) > 0. The vertical convergence of the wave energy flux is associated with the PKE maxima found over the convective region.
- However, over the central-eastern Pacific where deep cumulus convection is suppressed, there is strong equatorward fluxes of wave energy from the subtropics of the both hemispheres, causing horizontal convergence of wave energy flux in the equatorial upper troposphere. In this region a significant contribution from the barotropic conversion term is also noted. In contrast, poleward flows of wave energy from the equatorial region to higher latitudes is not observed.

b. Suggestions for future work

More extensive and focused examination of the structure and evolution of the MJO in terms of its dynamical properties in combination with the budget analysis of the energy equations are desired. Examinations of the properties and processes sensitive to the excitation and/or maintenance mechanisms are desired. Additional analysis of the air–sea interaction processes (e.g., Flatau et al. 1997; Wang and Xie 1998; Waliser et al. 1999) may result in a distinguishing conclusion.

1) Convection–Motion Coupling

The examination of the vertical profiles of the apparent heat source \( Q_1 \) and the apparent moisture sink \( Q_2 \) with the vertical structure of the MJO and its subcomponents will add information regarding the coupling between convection and large-scale motion. In the past, the knowledge of the mean vertical profiles of \( Q_1 \) and \( Q_2 \) yielded information about the prevailing cloud regimes of the western Pacific, the Atlantic trades and ITCZ, and other areas (e.g., Yanai et al. 1973; Nitta and Esbensen 1974; Thompson et al. 1979; Yanai and Tomita 1998). In the COARE IFA, Lin and Johnson (1996b) and Johnson and Lin (1997) found a variety of the vertical profiles corresponding to different convective regimes. Tung et al. (1999) further examined the basic modes of coupled \( Q_1 \) and \( Q_2 \) appearing under different large-scale conditions during the COARE IOP. Eventually, a holistic view of the coupling between large-scale motions and these convection regimes should be established.

2) Equatorial Waves as Components of MJO

Possible interference of westward moving Rossby and other waves with the eastward moving waves and cloud clusters should be studied:

(i) to investigate the mechanism of multiscale structure of the super cloud clusters. Chen et al. (1996), Takayabu et al. (1996), and Haertel and Johnson (1998) investigated the manner in which the cloud clusters were organized on temporal and spatial scales during TOGA COARE. They showed that, within the eastward propagating envelope marking the convectively active phase of the MJO, cloud clusters were frequently concentrated into westward propagating disturbances with a periodicity of ∼2 days.

(ii) to clarify the mechanism of wave energy emanation and accumulation in regions remote from the warm
The observed wave energy flux (thus group velocity) indicates the importance of Rossby waves and mixed Rossby–gravity waves in emitting and accumulating the energy generated in the warm pool region (Webster and Chang 1988).

3) Triggering

The two super cloud clusters studied in this paper originated near 75°E. This longitude is far west of the preferred longitudes of quasi-stationary convection in the eastern Indian Ocean, and, considering the earlier onset of wind and temperature perturbations, may suggest some triggering mechanisms coming from the west. As in the case of the super cloud cluster B studied here, convection may be excited by the overlapping of upper-level disturbances that are remnants of previous MJO events.

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