Observed Madden-Julian Oscillation in the Western Pacific Warm Pool:
A review

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Abstract

The convection in the warm pool exhibits multiscale variability. Besides its pronounced seasonal and interannual variability, it fluctuates significantly in the subsseasonal time scale and is strongly modulated by the Madden-Julian oscillation (MJO). This article reviews the MJO characteristics in the western Pacific warm pool, its role in the ocean-atmosphere interaction and its influence on the Australian monsoon and typhoon activity based mostly on observational studies.
1. Introduction

Sea surface temperature (SST) fluctuation is one of the most influential factors driving atmospheric convection. Among the many high SST regions, the warm pool in the equatorial Western Pacific not only contains the highest SST on Earth during the boreal winter but also harbors the most active convection. This region also releases the largest amount of diabatic heating due mostly to the latent heat release in the boreal winter. This is an important forcing influencing the planetary scale circulation.

The SST, convection, and circulation in the warm pool exhibit multiscale characteristics, ranging from diurnal to interdecadal variation. Among these different time scales, sub-seasonal time scale is one of the most significant variations that have been identified. Madden and Julian (1971, 1972) unexpectedly discovered quasi-periodic perturbation with a period of 40-50 days in the tropical Indian and Western Pacific Oceans. This phenomenon was named the Madden-Julian Oscillation (MJO). The MJO tends to reach its peak amplitude in the warm pool and dissipates quickly after leaving the warm pool. The MJO, which has been observed in many atmospheric and oceanic parameters, significantly modulates the convection and the diabatic heating in the warm pool. Its effect can subsequently have significant impacts on the low frequency variability in both global and regional circulation.

The MJO in the Western Pacific warm pool is one of the objectives of the Coupled Ocean-Atmosphere Response Experiment (COARE, Godfrey et al. 1998) carried out in the 1992/93 winter. Since then, many interesting characteristics of the MJO have been discovered. It would be worthwhile reviewing the research progress in the MJO ten years later after the completion of the COARE. This article reviews the observed characteristics of the MJO in the Western Pacific warm pool. Section 2 describes the general characteristics of the MJO, including the vertical and horizontal structure, the convection-circulation relationship, and the dependence of seasonality.
Features of different temporal and spatial scales embedded in the MJO are discussed in Section 3. The atmosphere-ocean interaction associated with the MJO is presented in Section 4. Section 5 describes the possible impacts of the MJO on the onset of the Australian summer monsoon and typhoon. This review focuses on the observational evidence. Theoretical and numerical aspects are not discussed. As shown later, the MJO is more active in the equatorial Western Pacific warm pool during the boreal winter than the summer. This review therefore focuses mostly on the phenomenon occurring during the boreal winter.

2. General characteristics

Madden and Julian (1971 and 1972) unexpectedly discovered significant spectral peaks around 40-50 days when exploring the characteristics of equatorial waves, such as the Kelvin and Yanai waves, by analyzing the circulation variables at tropical meteorological stations. Their cross-spectral analysis clearly identified 40-50-day signals (Figure 1) that originate in the Indian Ocean and move eastward across the Maritime Continent into the Western Pacific warm pool. The existence of this newly discovered feature was soon confirmed by several studies (e.g., Parker 1973). Similar signals were also found in cloud variations (e.g., Gruber 1974, Zangvil 1975). Cross-spectral analysis on different convection and circulation variables (e.g., Madden and Julian 1972, Hartmann and Gross 1988, Nishi 1989) indicated that the MJO is a large-scale circulation feature strongly coupled with the fluctuations in tropical convection. For example, the precipitation at tropical stations tends to precede the 850 hPa westerly by several days. The zonal winds in the upper and lower troposphere are highly coherent and generally out of phase. This reflects the dominance of the first baroclinic vertical structure in the tropics. These features can be seen in the schematic diagram shown in Figure 1.
Later studies found that the MJO is a broadband, with periods ranging from 20 to 80 days, and recurrent phenomenon in the Indian and Western Pacific (Madden and Julian, 1994) Oceans. Anderson et al. (1984) suggested that this feature is present 58% of the time, while Knutson et al. (1986) came up with an even higher 75%. The most frequently observed period is about 45 days (Madden and Julian 1994), which happens to be the average period. This recurrent phenomenon evidently accounts for a large portion of the total variability in both large-scale circulation and convection over the Indian and Western Pacific Oceans.

While there is no obvious change in the average period with seasons (Anderson et al. 1984), the seasonal variation in amplitude is significant (Madden 1986). Hartmann and Gross (1988) reported that in the Western Pacific warm pool the 40-50 day peaks in the 200 hPa zonal wind spectra are stronger and more prevalent in the boreal winter than in other seasons. In addition to the phase-lock into a seasonal cycle, the intraseasonal OLR variation exhibits large fluctuations in the climatologically convective regions, e.g. the Interhemispheric Convergence Zone (ITCZ), South Pacific Convergence Zone (SPCZ) and Asian monsoon regions. These characteristics have been noted in many studies (e.g., Madden 1994, Salby and Hendon 1994). Similar seasonality can be seen in Figure 2, which presents the ratio for the 20-100-day variance (hereafter referred to as intraseasonal variability, ISV) to the total variance of outgoing longwave radiation (OLR) for the boreal winter (November to April) and summer (May to October). Although the chosen band is broader than the aforementioned 20-80 days, the results are similar.

For this review, our discussion will focus on the ISV in the Western Pacific. During the boreal winter, the largest ISV in the Western Pacific is located mainly in the Southern Hemisphere over the warm pool and accounts for more than 25% of the total variance (Figure 2a). Intraseasonal fluctuations, however, prevail north of the
equator in the Western North Pacific during the boreal summer, leading to a weaker ISV in the Western Pacific warm pool (Figure 2b). Salby and Hendon (1994) found that the eastward-moving large-scale OLR signal amplifies from the boreal winter to the spring and decays thereafter when it migrates to the Northern Hemisphere. The Hovmöller plot of the intraseasonal and total OLR variances averaged between 120°E and the dateline shown in figure 3 essentially reproduces these characteristics. The close match between the seasonal migration of the total and 20-100-day variance is evident. The ISV south of the equator during the boreal winter is larger than its northern counterpart during the boreal summer. This asymmetry further indicates the importance of the MJO contribution to the ISV in the Western Pacific warm pool. Similar seasonal variations were also observed in the other variables, such as the zonal wind.

The intraseasonal OLR variation is dominated by a dipole pattern with extremes located in the Indian Ocean and the Western Pacific (e.g., Lau and Chan 1985, Zhu and Wang 1993, Hsu 1996). When the convection in the Indian Ocean is suppressed (active), its counterpart in the equatorial Western Pacific is often in the active (suppressed) phase. The dipole moves from the Indian Ocean along the equator across the Maritime Continent and often shifts southward while moving eastward in the Western Pacific (Wang and Rui 1990). The signal is amplified in the Indian Ocean, and often weakens slightly in the Maritime Continent, re-amplified in the Western Pacific, and stalls and weakens before reaching the dateline (Lau and Chan 1985, Wang and Rui 1990). Because of the dominance of this dipole pattern, convection flaring in the Maritime Continent and Western Pacific warm pool is often preceded by strong convective activity in the Indian Ocean.

The corresponding zonal scale and propagation of the MJO has been studied extensively using space-time spectral analysis. The convective signals near the
equator exhibit peaks between 35-95 days. This intraseasonal variance is contributed mostly by the eastward-propagating perturbations with a zonal scale equivalent to wavenumber 1 to 3 (Salby and Hendon 1994). Although these analyses were globally based, the signals mainly reflect the fluctuations associated with the aforementioned dipole. The same zonal scale and period were also found in the corresponding circulation (e.g., Gutzler and Madden 1989, Salby and Hendon 1994).

The MJO movement is a combination of eastward propagation and standing oscillation. Hsu et al. (1990) noted an eastward jump in the deep convection from the Indian Ocean to the tropical Western Pacific during the strong 1985/86 intraseasonal oscillation. A similar convection shift also occurred in the November/December 1981 event (Weickmann and Khalsa 1990). Cold surges preceding the eastward convection shift were observed in both cases. It was proposed that a cold surge triggering effect on tropical convection in the equatorial Western Pacific might lead to this sudden shift. However, long-term statistics to support this conjecture have not been established. Nevertheless, the discrete eastward movement of the MJO is a noted fact.

Weickmann and Khalsa (1990) reported the increasing accumulation of moisture in the lower troposphere in the tropical Western Pacific before the convection shift. Khalsa and Steiner (1988), based on satellite measurements from October 1981 to December 1985, showed the existence of stationary maximum centers of low-level precipitable water near 100°E and 150°E. These regions correspond to the maximum variance ratio shown in Figure 2a and are collocated with the climatological maximum moisture and SST regions. They are also the regions where the MJO tends to reach peak strength, as shown by Wang and Rui (1990). The tendency for the MJO to amplify in high moisture content and SST regions is due mainly to the positive feedback between the convection and large-scale circulation (Salby et al. 1994; Wang and Xie 1998). Zhang and Hendon (1997) suggested that this tendency yields the
impression of standing oscillation occurring in the equatorial Indian and Western Pacific Oceans.

The propagation of an intraseasonal OLR anomaly is shown in Figure 4 by vectors that point from each grid point toward the location where the lagged correlation is the largest five days later. This methodology, which is an efficient way to summarize the propagation tendency, has been applied in several studies (e.g., Lau and Chan 1985, 1986a; Hsu and Weng 2001). Consistent with the seasonal dependence of MJO amplitudes, the propagation is most prominent in the amplification regions where the moisture content in the lower troposphere and the SST are both high. During the boreal winter, the eastward propagation prevails south of the equator in the Western Pacific and Indian Oceans (figure 4a). The propagation speed is the largest in the Indian Ocean and becomes smaller towards the Western Pacific. The eastward component of the arrows becomes very small near and beyond the dateline, reflecting the stalling and dissipation of the system when moving farther eastward over the cooler equatorial Eastern Pacific. In contrast to the boreal winter, the eastward propagation near the equator is much less evident in the boreal summer (figure 4b). Instead, the most prominent features are the northward propagation in the Asian monsoon region and the Western North Pacific. The OLR anomalies used in Figure 4 were spectrally smoothed in space to retain only the large-scale convective system. When the non-smoothed data were used, the eastward propagation occurs mostly south of the Equator in the Indian Ocean and does not appear in the Western Pacific. This contrasting result is consistent with the dominance of planetary-scale perturbations (e.g., wave number 1-3) in the MJO as discussed above.

2. Multiscale variability

The MJO is a multiscale phenomenon including mesoscale, synoptic and large
scale features. Nakazawa (1988) was among the first to report these interesting characteristics. The convection system embedded in a MJO was found to be comprised of super cloud clusters (SCC) that exhibit a zonal scale of several thousand kilometers and moves eastward at a phase speed of about 10-15 m/s. Embedded in the SCC are mesoscale convective cells that are organized in westward-moving cloud clusters about 1000-2000 kilometers on a zonal scale. These cloud clusters have a 1-2 day lifetime. A case observed in the 1992/93 winter is shown in Figure 5 to illustrate these characteristics (Chen et al. 1996). The multiscale nature of the intraseasonal convective system has been studied extensively since then (e.g., Sui and Lau 1992, Lau et al. 1989, Hendon and Liebmann 1994). A schematic diagram proposed by Lau et al. (1989) is shown in Figure 6 to illustrate the spatial relationship between different scales of perturbations.

Significant fluctuations in the convective systems are accompanied by strong variations in atmospheric circulation in the tropics. As illustrated in the schematic diagram shown in Figure 6, the deep convection in the tropical Western Pacific is accompanied by low-level easterlies and upper-level westerlies near the equator to the east. This is associated with a pair of upper-level anticyclonic gyre and a pair of low-level cyclonic gyre to the west. In between the rotating gyres, upper-level easterlies and low-level westerlies exist. The cyclonic gyres were frequently observed to drift westward and poleward and developed into tropical storms. The aforementioned large-scale circulation pattern is often referred to as the Kelvin-Rossby wave packet forced by tropical heating (Gill 1980) because of its resemblance to the wave packet.

A low-level westerly in the Western Pacific warm pool often appears suddenly and increases to a significant strength in a short time. Wyrtki (1975) refers to this phenomenon as the westerly wind burst (WWB). The origin of the WWB has been an interesting subject because of its strong connection with the deep convection and the
ocean-atmosphere interaction. Lau et al. (1989) conducted a numerical study and suggested that this may be due to 1) a direct response to the passage of the eastward propagating main SCC heat source associated with the MJO and/or 2) zonal wind fluctuations on the equatorial side of the twin cyclones emanating from the main heat source.

However, Hendon and Liebmann (1994) found that the synoptic-scale disturbances in the 850 hPa zonal winds during the wet phase of the MJO move predominantly westward, instead of eastward, along the equator with a zonal wavelength of about 6000 km. Although the cross correlation between the 850-hPa zonal wind and the synoptically filtered OLR indicated an eastward propagation of the wind disturbances, it accounts for only 5% of the synoptic-scale wind variance. In the high-frequency mesoscale regime, they again found no significant cross correlation between the convection and the 850-hPa zonal winds. They subsequently concluded that the SCC is not a salient feature of the MJO and the role of the mesoscale mode in the MJO appears to be more of a consequence than a cause. These observations are not consistent with the hypothesis proposed by Lau et al. (1988), although the reliability of the tropical wind in the ECMWF data was questioned. Conversely, Houze et al. (2000) identified evidence, based on radar observations, suggesting a close relationship between the convection and strong westerly wind. It is likely that this relationship is strong in some cases, especially those with strong convection, but is not a dominant factor in the statistical sense.

Harrison and Vecchi (1997) investigated westerly wind events (WWE) in the tropical Pacific between 1986 and 1995 based on the 10-m winds from ECMWF. Eight types of westerly wind events were identified. Three of them occurred in the tropical Western Pacific warm pool. It was found that the near-equatorial WWE are sometimes, but not always, associated with cyclonic circulations on either (or both)
hemispheres. Many of the WWE were simply flows associated with zonal pressure
gradients and some of them were related to the cross-equatorial flow resulting from
the rapid increase in the meridional pressure gradient associated with cold-surge like
perturbations, e.g., Love (1985) and Chu (1988), and Chu and Frederick (1990).
Harrison and Vecchi (1997), however, did not evaluate the relationship between the
WWE and MJO. The above results suggest that the WWB (or WWE) can have
complicated origins. The MJO usually brings about the WWB in the Western Pacific.
However, the appearance of the latter is not necessarily related with the MJO all of
the time. The Origin the WWB remains an unsolved problem.

The multiscale nature is further confirmed by the observation taken during
COARE and the following studies, e.g., Chen et al. (1996), Johnson et al. (1999,
2001), and Houze et al. (2000). While both synoptic scale and mesoscale features
were observed embedded in two strong MJO events during the 1992/1993 winter
(Figures 5), there is no distinct scale separation between the convection and large-
scale disturbances during the intense convection periods (Chen et al. 1996). Houze et
al. (2000) studied the convection in two regions of the large-scale circulation
exhibiting Kelvin-Rossby wave characteristics, namely the westerly onset region
where the westerly meets the easterly in the lower troposphere and the strong westerly
region west of the westerly onset region. Distinct interactions between the SCC and
the mesoscale convection were identified in the two regions based on momentum
transport by the convection. The mesoscale inflow to the SCC transported easterly
momentum downward in the westerly onset region and westerly momentum
downward in the strong westerly region. The former would decelerate the low-level
westerly and therefore produce negative feedback to the background flow (e.g., the
WWB). Conversely, the latter would accelerate the low-level westerly and produce
positive feedback. The deceleration in the east and the acceleration in the west led to
enhancement of the low-level convergence to the west of the original convection. This phase shift between the low-level convergence and the deep convection caused the individual cloud cluster to move westward. The westward-propagating inertio-gravity waves, which exhibit similar characteristics, were suggested to explain the westward-moving 2-day cloud clusters (Chen et al. 1996).

Significant modulation of the atmospheric mixed layer at the intraseasonal time scale was observed during COARE (Johnson et al. 2001). The mean depth of the mixed boundary, 562m during the light-wind undisturbed period, decreased to 466m during the heavy-rain period of the WWB embedded in the MJO, and then increased to 629m in the later stages of the WWB. Johnson et al. (1999) reported observations of the trimodal characteristics of tropical convection in COARE. In addition to cumulus and deep cumulonimbus, cumulus congestus, which was largely ignored in the past, was frequently observed and contributed over one-quarter of the total convective rainfall. The three cloud types varied significantly during COARE on the intraseasonal time scale. The shallower congestuses contributed to moisten and precondition the atmosphere for deep convection, while the deeper congestuses contributed a significant amount of total tropical rainfall. It was suggested that both types of congestus were likely to produce many mid-level clouds and modulated the radiative forcing to the large-scale circulation in the tropical Western Pacific.

The passage of the convective system embedded in the MJO undoubtedly modulates the mixed layer and clouds, which may in turn produce feedback to the large-scale system through various mechanisms. It is however not clear so far how strong the interaction is and how it works, if it exists. The scale interaction may enhance or weaken the MJO. However, whether it is essential for the genesis and maintenance of the ISO remains to be seen.
3. Ocean-atmosphere interaction

Intraseasonal variability was observed not only in the SST (e.g., Kawamura 1988; Zhang and McPhaden 1995; Zhang 1996, 1997; Hendon and Glick 1997; Woolnough et al. 2000), but also in the latent and sensible heat fluxes and radiation (e.g., Jones et al. 1998) at ocean surface, which are key variables reflecting the ocean-atmosphere interaction. It has been found that the SST fluctuation on the intraseasonal time scale can largely be explained by surface fluxes (e.g., Zhang 1996; Hendon and Glick 1997; Shinoda et al. 1998). A schematic diagram adapted from Flatau et al. (1997) is shown in Figure 7 to illustrate the air-sea convective intraseasonal interaction (ASCII), which is basically the wave-conditional instability of the second kind (CISK) modified by the ocean surface heat fluxes. To the east of the deep convection embedded in an eastward-moving MJO, the SST rises due to the excessive shortwave radiation reaching the ocean surface and weaker evaporation in the less cloudy and weak wind condition. Upon the arrival of the deep convection, the downwelling shortwave radiation weakens and the SST starts dropping. The strong westerly to the west of the deep convection encourages strong evaporation and the turbulent mixing in the ocean mixed layer. As a result, the SST drops significantly (in some cases, by 1 °C, e.g., Lukas and Lindstrom 1991). The moisture is then transported into the deep convection, condenses during ascent in the convection core, and releases latent heat to sustain the deep convection, which in turn drives the large-scale circulation to provide more moisture and heating. This ocean-atmosphere interaction through heat flux exchange and wind stress has been confirmed by observations during COARE (e.g., Weller and Anderson 1996, Lin and Johnson 1996, Lau and Sui 1997, Sui et al. 1997, Chou et al. 2000).

The positive feedback between the large-scale circulation and convection is the key mechanism that sustains the MJO. As discussed above, it has been demonstrated
by several studies (e.g., Salby et al. 1994, and Wang and Xie 1998) that this process is more active in the high SST and moisture content regions. The Western Pacific warm pool is one of the highest SST regions and is therefore a region for the MJO amplification. While the above mechanism may partially explain the maintenance and amplification of the MJO in the high SST regions, whether the ocean-atmosphere interaction is a necessary condition for the existence of the ISO is not yet clear. Slingo et al. (1996) examined the performance of the atmosphere general circulation model (AGCM) in simulating the MJO and found poor results in most models. The reasons are varied and complex. Lack of ocean-atmosphere interaction is one of the possible candidates. Studies (e.g., Walizer et al. 1999, Hendon 2000) using coupled ocean-atmosphere models however yield mixed results. This could be due to the inaccurate representation of the ocean-atmosphere interaction in the coupled models. On the other hand, it has also been demonstrated (e.g., Maloney and Hartmann 1999) that the performance can be improved simply by adjusting the physical parameterization schemes, e.g., the convection scheme, in an AGCM. It is likely that the MJO is an intrinsic atmospheric mode enhanced by the ocean-atmosphere interaction.

Besides the SST, intraseasonal fluctuations were also observed in sea level (Enfield 1987; Chelton and Schlax 1996), thermocline (Lukas and Lindstrom 1991, Kessler et al. 1995; Zhang 1997), ocean currents and ocean waves (Johnson and McPhaden 1993a,b; Kessler and McPhaden 1995, Hendon et al. 1998). Wyrtki (1975) suggested that the WWB in the tropical Western Pacific could excite the oceanic Kelvin waves that propagate eastward across the Pacific to the equatorial Eastern Pacific after 3 months to induce the onset of El Niño. Since then, many studies have proposed that the MJO in the Western Pacific may have a triggering effect on the El Niño (Lau and Chan 1986b, 1988; Kessler et al. 1995, Kessler and Kleeman 2000). This hypothesis has been supported by observations of active intraseasonal variability

The triggering mechanism works as follows. The strong westerly wind in the tropical western Pacific excites downwelling Kelvin wave (Kessler et al. 1995, Hendon et al., 1998, figure 2) as shown in Figure 8, which reduces the upwelling in the equatorial Eastern Pacific, advects the warm Western Pacific water eastward, and results in the SST increase in the equatorial Central Pacific. The sea surface cooling in the equatorial Western Pacific, due to evaporative cooling and turbulent mixing, and the warming in the Central Pacific result in the eastward shift of the convective activity, which in turn leads to the eastward shift of the westerly wind anomalies. As a result, the sea surface warming occurs farther to the east because of this positive feedback between the SST and the surface wind.

While the atmospheric intraseasonal variation have a strong influence on the ocean, one may wonder whether intraseasonal variability would be enhanced over a warmer sea surface on the interannual time scale. In a study on the interannual variation of 30-90-day variance, Hendon et al. (1999) concluded that the interannual variation is primarily associated with changes in the number of discrete MJO events (Figure 9a) and with changes in the intensity of intraseasonal convection in the Indian Ocean and the Western Pacific. While the location of major MJO activity tends to shift eastward during the El Niño, the overall level of MJO activity was uncorrelated with El Niño (figure 9b), except during exceptional warm events (e.g., 1982/83 and 1997/98) when the MJO activity is diminished. It was also shown that the MJO activity level is weakly related to the SST anomaly in the equatorial Indian and Western Pacific Oceans. In other words, much of year-to-year variability of the MJO is internally generated.
4. Relationship with Australian monsoon and typhoon

As discussed before, the location of major MJO activity migrates meridionally with the climatological convection regions. Several studies reported that the Australian summer monsoon is modulated by the eastward-propagating MJO. The wet-westerly phase of the Australian summer monsoon is often preceded by the MJO propagating eastward from the Eastern Indian Ocean into Northern Australia, and is followed by the MJO propagating away from Australia and into the Western Pacific (Holland 1986, Hendon and Liebmann 1990a). Hendon and Liebmann (1990b) also found that the onset of the Australian summer monsoon tends to coincide with the passage of a MJO. Drosdowsky (1996) however concluded that there is no clear relationship between westerly and rainfall in the Australian summer monsoon. In addition, no dominant time scales were found in the length of the convection-active periods or in the recurrence time between active phases. On the other hand, his conclusions do not exclude the influence of the MJO. Among the 59 events shown in Figure 12 of Drosdowsky (1996) 29 events exhibited recurrence time between 30-60 days, accounting for almost half of all events. This contrasting result suggested that the MJO might have impacts on the Australian summer monsoon but is not necessarily the dominant factor.

Multiscale convection tends to be more active during the wet phase of the MJO. Clustering of tropical cyclone and depression has been observed in several studies. Nakazawa (1986) reported that tropical cyclones in the Western North Pacific tended to occur in the negative phase of the intraseasonal OLR anomalies during the 1979 summer. Sui and Lau (1992) noted the tendency of the cyclone formation occurring upon the arrival of a SCC embedded in the MJO. Four cyclonic gyres developed into tropical cyclones during COARE (Chen et al. 1996). In a statistical
study, Liebmann et al. (1994) concluded that cyclones preferentially occurred in the wet phase of the MJO (figure 10) and clustered around the low-level cyclonic vorticity and divergence anomalies that appear poleward and westward of the large-scale convective anomaly along the equator. Hall et al. (2001) found similar results in the Australian region. Liebmann et al. (1994) however also found that this association with the convective region is not unique to the MJO and a depression is no more likely to develop into a cyclone in the wet phase than in the dry phase. The influence of the MJO is important not because it changes the characteristics of tropical cyclogenesis; instead, it is because of its modulation and clustering effect on the tropical cyclone activity.

5. Conclusion

The MJO has received so much attention since its discovery simply because its tremendous impacts on the low frequency variability of the atmospheric general circulation. It is almost compatible with the celebrated phenomena such as ENSO, sudden warming, Pacific Decadal Oscillation, Arctic Oscillation, and so on. Studies have been carried out based on field experiments such as COARE, data analysis, numerical simulation, and tropical wave theory. We have been able to identify many of its interesting characteristics and proposed many theories to explain its structure and evolution.

During the course of studies, more detailed and intriguing features associated with the MJO were discovered. Some of them have been satisfactorily explained. Some of them stirred up more discussions and controversies. Its possible effect in inducing the ENSO is one. Although some studies indicate that the significant statistics is not found, there have been indications of the strong MJO activity preceding the occurrence of El Niño ad La Niña. The shift of convection from the
Eastern Indian Ocean to the warm pool is evident but is hardly studied. There are still much room for the study on its relationship with the onset of the Australian summer monsoon and typhoon. Studies are still needed to understand its relationship with the embedded small-scale convection and boundary layer processes. One important area that has been hardly studied is the detailed cloud structure and the role of cloud-radiative feedback.

More importantly, most of the present state-of-art GCMs are incapable of simulating the MJO (Slingo et al. 1996). Different reasons, including convection schemes, boundary-layer process, ocean-atmosphere interaction, have been proposed. However, there are no easy answers. Different models might need different remedies. This problem reveals that we are still far from really understanding the MJO and that there are still many mysterious nature of the MJO needed to be explored and explained.
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Figure caption

Figure 1. Schematic diagram to illustrate the characteristics of the large-scale circulation and convection associated with an eastward-moving MJO. Time is progressive from top to bottom. Pressure perturbations are plotted at the bottom of each chart with negative anomalies shaded. The circulation cells are based on the mean zonal wind disturbance. Regions of enhanced large-scale convection are indicated schematically by the cumulus and cumulonimbus clouds. The negative tropopause height is indicated at the top of each chart (adapted from Madden and Julian 1972).

Figure 2. The total OLR variance and the ratio of the 20-100-day filtered OLR variance to the total variance for (a) winter (November-April) and (b) summer (May-October). The total variance is plotted by contours and the interval is 500 W²/m⁴. The shading indicates the ratio and the interval is 5 percent as indicated at the bottom of the figure.

Figure 3. Hovmöller diagram of the total and 20-100-day filtered OLR variance to illustrate the seasonal evolution of the convection. The former is contoured for every 200 W²/ m⁴ and the latter is shown in shading for every 50 W²/ m⁴.

Figure 4. Tendencies of intraseasonal OLR anomaly propagation derived from 5 day lagged correlation maps for (a) winter (November to April) and (b) summer (May to October). Arrows indicate the direction and distance for an anomaly propagating from the base point to another location in 5 days. Contours indicate the 5-day lag correlation coefficients, which are not shown unless larger than 0.7 are shaded. Shading interval is 0.05.

Figure 5. Hovmöller diagram of convective activity in the tropical eastern Indian and western Pacific Oceans during December 1992 when MJO events were observed. Multiscale perturbations in different propagation direction are evident (adapted
from Chen et al. 1996).

Figure 6. Schematic diagram to illustrate the hierarchy structure of circulation and convection in a MJO over the tropical Pacific Ocean (adapted from Lau et al. 1989).

Figure 7. Schematic diagram to illustrate the air-sea convective intraseasonal interaction (ASCII) mechanism (adapted from Flatau et al. 1997).

Figure 8. Hovmöller diagram indicating the relationship between the fluctuation in the depth of 20C SST (Z20) and the wind stress. The Z20 pattern was derived from empirical orthogonal function analysis and is shown in shading. Wind stress is shown by contour. It is clearly indicated that the westerly (easterly) wind stress in the tropical western Pacific leads the positive (negative) depth anomalies in the tropical eastern Pacific (adapted from Hendon et al. 1998).

Figure 9. (a) Fluctuations of number of MJO events per year and averaged period, and (b) variations of overall variance of zonal wind at 200 and 850 hPa and OLR from 1974 to 1997. Nino3 index is also shown in (b) for comparison (adapted from Hendon 1999).

Figure 10. Number of typhoon, tropical storm, and tropical depression in the tropical Pacific and Indian Ocean during the wet (negative OLR, 200 hPa divergence, and 850 hPa convergence anomalies) and dry (positive OLR, 200 hPa convergence, and 850 hPa divergence anomalies) phases of the MJO along (a) 12°N and (b) 12°S (adapted from Liebmann et al. 1994).
Figure 1
Figure 2a

Figure 2b
Figure 4a

OLR 20–100dy–filtered spatial_filtered winter lag +5dy vectors

Figure 4b

OLR 20–100dy–filtered spatial_filtered summer lag +5dy vectors
Hierarchy structure of intraseasonal oscillations 
over the Tropical Pacific Ocean

Intraseasonal (30-60 days) Oscillations 
(entire tropics)

Deep convection regime 
(western and central Pacific)

Super cloud clusters 
(2-4000 km)

westward moving 
cloud clusters 
(1000-2000 km)

eastward moving major 
precipitation pattern 
(~2000 km)

continued eastward 
moving circulation

Figure 6
Figure 7
Figure 8
Figure 9a

Figure 9b
Figure 10