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An Overview of the Madden–Julian Oscillation and Its Relation to Monsoon and Mid–Latitude Circulation

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ABSTRACT

In the past decade there has been extensive research into tropical intraseasonal variability, one of the major components of the low frequency variability of the general atmospheric circulation. This paper briefly reviews the state–of–the–art in this research area: the nature of the Madden–Julian Oscillation, its relation to monsoonal and extratropical circulations, and the current theoretical understandings.

I. INTRODUCTION

Variations with periods longer than the deterministic predictability time scale of roughly 10 days are commonly referred to as low–frequency variations (Ghil and Mo, 1991a) or short–term climate variations. The low–frequency variations of tropical atmospheric circulation exhibit three major components–interannual, annual, and intraseasonal. The interannual variability is characterized by the E1 Nino–Southern Oscillation (ENSO) which has its strongest signals in the eastern and central tropical Pacific. The annual variation is characterized by monsoons and has its strongest signals in the South Asia–North Australia sector of the tropics. The intraseasonal variability is characterized by 40–50 day oscillation first discovered by Madden and Julian (1971), which has strongest signals in the tropical Indian and western Pacific Oceans.

The intraseasonal time scale ranges from 10 to 90 days. The atmospheric motion exhibits a significant portion of energy on this time scale. The outgoing longwave radiation (OLR) data, which provide an indirect measurement of the deep convection and rainfall over the global tropics, were used to estimate the variances on different time scales (Lau and Chan, 1988). The root–mean–square amplitude of annual variation in OLR averaged over global tropics between 40°N and 40°S is 15 W m^{-2}, while the amplitudes of high frequency (mainly 1–5 days), intraseasonal, and interannual (2–6 years) variabilities are approximately 18, 10, and 6 W m^{-2}, respectively. Over the tropical Indian and western Pacific, the intraseasonal variation carries about a quarter to one third of the total variances.

Within the broad frequency band of intraseasonal variation, the 40–50 day or Madden–Julian Oscillation (MJO) is most prominent among others. This paper presents a brief review of the synoptic and dynamic aspects of MJO. In the interest of brevity, no attempt will be made to describe the complete history of the research and summarize all of the current research results in this area. Our attention is focused on the observed features of MJO (Section 2), its relation to monsoon variations and the intraseasonal teleconnection between the tropics and midlatitudes (Section 3), and our current theoretical understanding of the physics of MJO (Section 4). Our discussion will only pertain to atmospheric circulation.

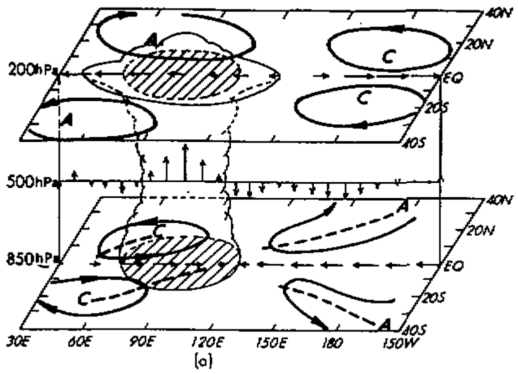

Fig.1. Schematic depiction of the characteristic structure of the Madden–Julian waves. The shaded regions correspond to areas where pentad mean OLR anomalies are less than $-7.5\ \mathrm{W\ m^{-2}}$. The circulation cells highlight characteristic wind anomalies associated with the convection anomalies (From Rui and Wang, 1990).

Detailed discussion of the air–sea interaction on intraseasonal time scale and its relation to ENSO are beyond the scope of the present review.

II. OBSERVED MADDEN–JULIAN OSCILLATION

Madden and Julian (1972) used cross spectral analysis of station wind data and revealed a systematic eastward propagation of the oscillation mode. They conjectured that the 40–50 day oscillation results from planetary–scale overturning cells moving eastward around the globe in the tropospheric equatorial plane. A renewed surge of interest developed in the early 1980s when OLR data and FGGE (First GARP Global Experiment, where GARP stands for Global Atmospheric Research Program) data became available. The existence of the MJO was successively confirmed in terms of velocity potential and winds (e,g., Lorenc, 1984; Krishnamurti et al., 1985; Knutson et al., 1986) and OLR (e. g., Weickmann et al., 1985; Lau and Chan, 1985, 1986; Murakami et al., 1986). These studies used various statistical (correlative, cross–spectral, and principal component) analyses and documented important mean behavior of MJO. For instance, the oscillation covers wide range of periods ranging from 30 to 60 days and has a zonal circulation scale of wavenumber one or two; the oscillation exhibits maximum strength in the deep tropics near the equator and affects strongly the transient fluctuation of the Pacific Walker cell and the South Pacific convergence zone (SPCZ). In the tropics, zonal wind fluctuations in the upper troposphere are nearly out of phase with those in the lower troposphere, while poleward of about 20° there is no pronounced phase shift between levels (Knutson and Weickmann, 1987).

From a dynamical viewpoint, however, it is more meaningful to identify and diagnose intraseasonal convection and circulation anomalies as individual dynamic systems (Madden–Julian mode). The movement seen directly from individual events is more realistic and complex. The eastward–moving intraseasonal convection anomalies exhibit three major paths: (1) eastward along the equator from Africa to the mid–Pacific, (2) first eastward along the equator, then turning at the maritime continent either toward the northwestern Pacific (often in boreal summer) or toward the southwestern Pacific (often in austral summer), and

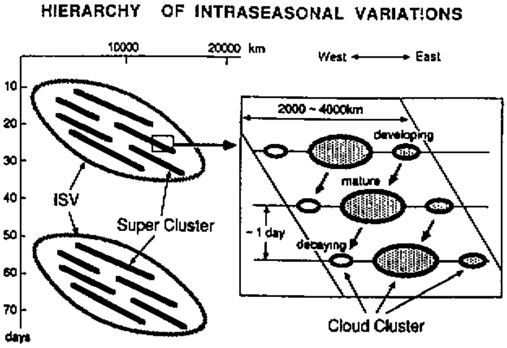

Fig.2. Schematic diagram for the hierarchy of Madden–Julian oscillation mode (from Nakazawa, 1988).

(3) the main anomaly moves eastward along the equator with splitting center(s) moving northward over the Indian and / or western Pacific Oceans (Wang and Rui, 1990a). In addition to dominant eastward propagation, evidences of westward migration were also presented (Chen and Xie, 1988; Lu et al., 1991), but the spatial distribution is limited and the strength is significantly weaker compared to eastward propagation.

The Madden–Julian mode displays remarkable longitudinal variations. Using sea–level pressure data, Madden and Julian (1972) noted eastward propagating waves having the largest amplitudes in the western Pacific. Study of the coherent variations of the OLR and circulation revealed coherent principal emperical orthogonal function (EOF) patterns of eastward propagating divergent and rotational waves in association with anomalous convective activity (Knutson et al., 1986; Knutson and Weickmann, 1987). In the Eastern Hemisphere the upper and lower circulation anomalies couple with convection and slowly move eastward at a speed of about 5 m s^{-1}, whereas in the Western Hemisphere only upper tropospheric circulation anomalies are evident which travel at a speed about three times faster (16 m s^{-1}) after they decouple from convection anomalies. On the basis of a composition of 36 individual eastward–moving events, Rui and Wang (1990) proposed a longitude–dependent four–stage life cycle: initiation over the equatorial Africa, rapid intensification when passing through the Indian Ocean, mature evolution characterized by a weakening in the maritime continent and re–development over the western Pacific, and dissipation near the dateline in the moderate events or emanation from the equator toward North America and southeastern Pacific in the strong events. The emanation from east of the dateline to North America is consistent with the findings of Lau and Phillips (1986). The swift shift from Indian Ocean to western Pacific warm pools is also seen in a case study of Weickmann et al. (1990). The equatorial Indian Ocean and the western Pacific monssoon trough and SPCZ are preferred geographic locations for the development of Madden–Julian mode, while the maritime continent and central Pacific are regions of dissipation (Wang and Rui, 1990a).

Documentation of the structure of the oscillation mode is no doubt of central importance for understanding the dynamics of MJO. Using composite near–equatorial station data, Madden (1986) showed that eastward–moving convection anomalies force a Kelvin–like wave to the east and anticyclonic Rossby–like waves to the west with dominant anticyclonic eddies occurring in the summer hemisphere. On the other hand, using EOF analysis of National Meteorological Center (NMC) wind and OLR data, Knutson and Weickmann (1987) found anomalous upper tropospheric cyclones to the east of the convection and anticyclones alongside or west of the convection with prominent rotational flow patterns in the winter hemisphere. By analysis of the six strongest eastward–moving events during 1979–1985 using European Center for Medium–range Weather Forecast (ECMWF) wind and OLR data, Rui and Wang (1990) presented a schematic depiction of the characteristic structure in which the equatorial upper– (lower–) level easterly (westerly) anomalies are coupled with equatorial convection anomalies and with twin anomalous anticyclonic (cyclonic) circulation anomalies (Fig.1). The wind anomalies generally lag convection anomalies in developing phase, but nearly overlap in late mature phase and slightly lead the convection anomalies in dissipation phase. As convection ceases in the central Pacific, the upper–level twin cyclonic cells associated with westerly anomalies in front of the convection can penetrate into eastern Pacific while the low–level wind anomalies die out east of the dateline. This quasi–symmetric structure shares similarities with the results of Knutson and Weickmann (1987) and Murakami (1988).

Another interesting feature is the multi–scale structure. Nakazawa (1988) showed that the planetary–scale Madden–Julian mode in the equatorial region is composed of several super clusters which have a horizontal scale of several thousand kilometers and move eastward. Each super cluster, however, consists of a group of westward–propagating cloud clusters which have an individual life span of 1–2 days. The eastward movement of super clusters is in a discrete manner due to the successive formation of a new cluster which propagates westward and becomes either damped or moves away from the equatorial regions (Fig.2). This finding raised a fundamental question regarding the manner in which the low–frequency component of the tropical atmosphere actually varies.

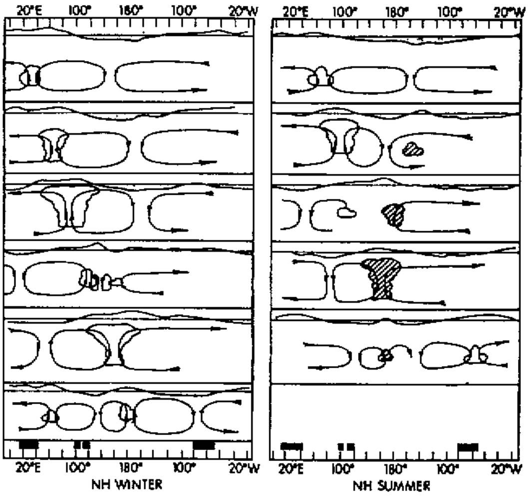

Fig.3. Schematic models of the Madden–Julian Oscillation projected onto the equatorial zonal plane for (a) boreal winter, and (b) boreal summer. The curves above the convection anomalies are velocity potential anomalies at 200 hPa with positive denoting divergence. The wind plotted represents the divergent component (From Zhu and Wang, 1991).

In addition to the transient wave character, the MJO also has a pronounced regional standing component. An out–of–phase relationship between the convection anomalies over the maritime continent and the central Pacific was noted by Lau and Chan (1985, 1986) who described this dipole pattern as a seesaw and interpreted it as resulting from systematic eastward propagation of equatorial waves. Zhu and Wang (1991) documented the prominent convection seesaw that exists between two permanent action centers of intraseasonal convectivity over the tropical Indian Ocean and western Pacific. When convection undergoes a major intensification over the Indian Ocean, the western Pacific usually experiences an abnormal dry condition; whereeas when convection develops over the tropical western Pacific, the convection in the Indian Ocean is often suppressed. The formation of the seesaw oscillation is season–dependent. Figs. 3a and 3b depict schematically large–scale processes of the intraseasonal oscillation which emphasize that the seesaw oscillation is an intrinsic feature of the transient wave activities. The boreal winter process (Fig.3a) is merely a modification of Madden–Julian's (1972) model with stresses on the weakening and / or splitting occurring over the maritime continent and redevelopment over the western Pacific (often in SPCZ). In the boreal summer model (Fig.3b), the regional seesaw results from a time–lagged development of two transient systems: the western system originates in the Indian Ocean and moves eastward and / or northward while the eastern system develops in the western Pacific and moves westward and / or northward. Both of them intimately interact with monsoon circulations. The common feature for both the winter and summer models is the eastward travel of the wavenumber–one divergence–wave around globe in 40–50 days. Why the seeaw has two poles residing both sides of the maritime continent needs to be further addressed.

There exists a clear annual variation in MJO. Although it occurs throughout the seasonal cycle without systematic variation in periodicity (Anderson et al., 1984), the locations of maximum OLR variability and its extratropical response does exhibit seasonality (Knutson and Weickmann, 1987). Madden (1986) found that the zonal wind anomaly in the 40–50 day time–scale exceeds that in adjacent lower and higher frequency bands by the largest amount during December through February. The events of eastward propagation not only take place more frquently but also attain stronger intensity during boreal winter, whereas the northward propagation events predominantly occur in boreal summer from May to October (Wang and Rui, 1990a).

III. OSCILLATIONS IN MONSOON DOMAINS AND INTRASEASONAL TELECONNECTION

1. *Oscillation in Summer Monsoon Domains*

The South Asia summer monsoon, one of the earth's most energetic weather events, influences the global atmospheric circulation. Yasunari (1979, 1980, 1981) first revealed a systematic northward propagation in cloudiness and wind fluctuation from the equator to the Himalayan region, exhibiting a near–steady meridional propagation of a train of zonally oriented troughs and ridges. The troughs are associated with rising motions and clouds while the ridges are essentially cloud–free. The meridional propagation of troughs and ridges on this time scale in the lower troposphere was further described by Krishnamurti and Subrahmanyan (1982). The phenomena of onset, active, and break monsoon appear to be related to the passage of these low frequency systems. The meridional scale of this mode is around 3000 km, and its phase speed is about 0.8 latitude per day. Julian and Madden (1981) postulated that this Yasunari mode is a portion of the global equatorial MJO. The relationship between the divergent wave (or fluctuation in Walker circulation associated with MJO) and smaller–scale divergence bands (inactive spells of Indian monsoon) were further confirmed (Lorenc, 1984; Krishnamurti and Gadgil, 1985; Murakami and Nakazawa, 1985). There is abundant evidence that demonstrates the existence of the oscillation with a period of 40 days in other monsoon regions such as the northwestern tropical Pacific (Chen and Murakami, 1988), East Asia (Chen and Jin, 1984), Australian monsoon region (Hendon and Liebmann, 1990), and the Southern Hemisphere tropics (Vincent and Sperling, 1991). In the rest of this subsection, a more detailed discussion will be devoted to the first two regions.

When the summer monsoon advances northward in East Asia, its leading zone, i.e., the monsoon rain belt correspondingly moves from low latitude to midlatitudes. In this process, as does the monsoonal air low, the monsoon rainfall belt undergoes three stages of relative standing in South China (~23°N), the Yangtze River Valley (~30°N), and North China (~40°N) which are linked by two stages of abrupt northward jump (Fig.4). In late spring the rainbelt is located in South China, while North China and the Yangtze River Valley are dry. This is the first standing stage of the major rainbelt (often referred to as the presummer rainy season) that normally continues until mid–June when it suddenly shifts to the valley of the Yangtze River. This northward "jump" is followed by a second standing stage corresponding to the so–called Plum Rains (Meiyu). By mid–July, the rain belt rapidly moves to North China and in August moves further to Northeast China, reaching the northernmost position of summer monsoon rainfalls. In late August the rainy season of North China ends and the major rain belt suddenly retreats back to South China, leaving most of the eastern part of China dominated by a dry spell in autumn. Note that from mid–July to September, South China experiences a second rainy season. There is relatively dry period of about one month between the presummer and the second rainy season. The latter is caused mainly by typhoons, monsoon troughs and other tropical disturbances. Around the beginning of October, the rainy season over Southeast China nearly ends and rapidly makes way for winter monsoon.

K. M. Lau et al. (1988) realized that the above–mentioned annual march of the rain belt in East China appears to be modulated by a primary 40–day oscillation with a 20–day oscillation superimposed at the later stage (Fig.5). As a matter of fact, the wet spell in the first cycle of low frequency oscillation broadly corresponds to the presummer rainy season over South China and / or the Meiyu period over the Valley of the Yangtze River whereas the second wet spell corresponds to the rainy season in North China. On the basis of analysis of the power spectrum of precipitation for 15 stations in China during June–August, 1979, this association between the 40–day oscillation and precipitation in China has been further documented (Zhou et al., 1986).

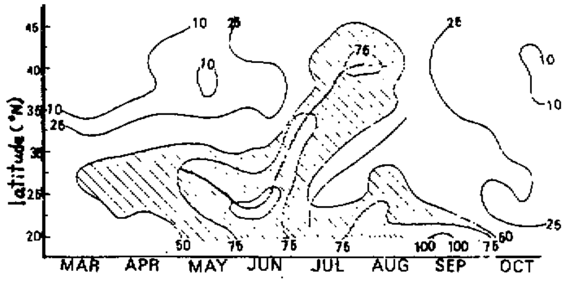

Fig.4. The temporal and spatial variation of mean 10—day precipitation amount in the eastern China averaged for 1950–1979 (Guo and Wang, 1981). Unit: mm.

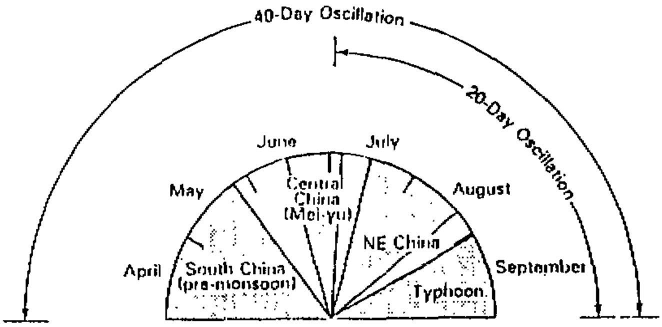

Fig.5. Schematic diagram showing the relationship between the major rainfall regimes in East Asia and the low frequency oscillations. The area of the shaded portion is proportional to the duration of the rainfall regimes. The gaps in between the rainfall regimes indicate transition periods. Periods during which the 40–day and the 20–day oscillations are found are indicated (from K.–M. Lau et al., 1988).

The 30–50 day oscillation in East Asia and the western Pacific is characterized by a large–scale break–active cycle of monsoon as well as corresponding variations in the intensities of monsoon meridional cell and subtropical high. Chen and Ke (1983) made a detailed analysis of the East Asian monsoon by use of a composite method. The variations of the 850 hPa wind pattern may be identified as consisting of four stages, each of which spans about 10 days. In short, the 30–50 day oscillation shows up in an alternation of northward advance of the monsoon trough as the equatorial westerlies enhance and move northward, and the southward withdrawal and dissipation of the monsoon trough as the equatorial westerlies weaken.

The 30–50 day oscillation in East Asia propagates eastward and northward, basically similar to the situation in the South Asian monsoon region. He and Yang (1990) revealed that in the East Asian monsoon region there is a confluent belt of the meridional propagation of low frequency meridional wind at low levels while a diffluent belt of the meridional propagation of low–frequency zonal wind at higher leves. These belts are located around 20–30°N. The properties of low frequency systems on either side of this belt differ remarkably. Therefore, the diffluent (or confluent) belt may be taken as a dividing zone between the tropical and midlatitude low frequency oscillation.

2. *Intraseasonal Teleconnection*

Using OLR data, Weickmann (1983) revealed a close association between intraseasonal extension and contraction of the time mean midlatitude jet and tropical cloudiness fluctuation. Anderson and Rosen (1983) detected oscillation in total atmospheric relative angular momentum. Both works indicate the global influence of MJO. Lau and Phillips (1986) showed that the time evolution of the Eurasia and Pacific–North America wave trains is coherent with that of tropical dipolar convection. While the extratropical anomalies occur at fixed geographic locations along the entire latitude circle, tropical convection is restricted to the Indian Ocean / western Pacific region. On the other hand, Liebmann and Hartmann (1984) emphasized a predominant energy propagation from midlatitude to tropics on an intraseasonal time–scale, especially in the eastern Pacific. Magana and Yanai (1991) showed that intraseasonal variabilities of the mid–Pacific trough and South Asia and Mexico anticyclones are coherent with convective activities in the Indonesia–western Pacific region and in ITCZ over central America. The mid–Pacific trough acts as a two–way link between the tropics and midlatitude: on the one hand, it plays dominant role in northward transport of westerly momentum; on the other hand, the westerlies that develop southeast of the trough form an equatorial westerly duct through which wave energy can propagate into tropics from midlatitude. Schubert and Park (1991) suggested that the convection in western North Pacific has a strong modifying influence on midlatitude low–frequency variability via changing the upper tropospheric zonal winds, whereas Chen and Yen (1991) showed that the tropical–midlatitude interaction is through the eastward propagation of intraseasonal velocity potential oscillation.

Nitta (1986) and Huang and Li (1988) stressed the teleconnection effect of the convective activity over the South China Sea and around the Philippines on the anomalous behavior of the general circulation in East Asia and Japan through the propagation of a stationary wave train. The propagation of this wave train has immediate downstream effect on the activity of the subtropical high over the northwestern Pacific. The subtropical high often shifts northward and the summer rainfalls are below–normal in such regions as the Yangtze River Valley, Huaihe River and South of Japan. Huang and Sun (1990) further indicated that this teleconnection pattern is closely related to the water temperature in the surface layer and the subsurface layer between 50 m and 300 m of the western Pacific warm pool. When the water temperature in this region is above–normal, i.e., the warm water is accumulated in the tropical western Pacific, the convective activity is intensified over the region from the Indo–China Peninsula to Philippines.

The thermal condition at the land surface, in particular the Tibetan Plateau, also exerts a significant effect on the general circulation at mid–and high latitudes through the teleconnection pattern. If the heat source over the plateau gets enhanced, the Tibetan high in turn will intensify. Then the trough north of this high will deepen, thus causing a sequence of downstream developing events (Gambo and Kudo, 1983). This teleconnection may be used to predict the drought / flood in North China and the northern Japan, based on the anomalous changes in heating or snow cover fields over the Tibetan Plateau.

IV. DYNAMICS OF THE MADDEN–JULIAN OSCILLATION

1. *Understandings from General Circulation Model (GCM) Simulation*

Many essential features of MJO have been simulated using various versions of GCMs. The ability of an aqua–planet GCM (e.g., Hayashi and Sumi, 1986; Swibank et al., 1988) and GCM with axially symmetric climatology (e.g., Lau and Lau, 1986) in simulating reasonable MJO phenomenon demonstrates that neither land–sea thermal contrasts nor zonal asymmetry is necessary to explain the basic mechanism of the oscillation (Emanuel, 1987). These numerical simulations demonstrate that the model counterparts of the MJO are caused by eastward movements of planetary scale (wavenumber one or two) circulation modes which are primarily confined to 15° latitude from the equator and vertically penetrate the entire troposphere with a baroclinic structure. The simulated phase propagation (10–20 ms^{-1}) in many experiments tends to be faster than observed, however models with higher resolution may yield comparable phase speed (Hayashi and Golder, 1986). Most of numerical studies using idealized GCM models suggest that condensational heating interacting with dynamic processes is responsible for the generation and maintenance of these long–lasting planetary scale modes which resemble either a Kelvin wave–CISK (conditional instability of second kind) mode (Lau and Peng, 1987; Swinbank et. al., 1987; N.C. Lau et al., 1988), or a coupled Kelvin–Rossby mode (Hayashi and Sumi, 1986; Hayashi and Golder, 1988; Hendon 1988).

The slow northward propagation on a time scale of 20–40 days has been simulated in terms of zonally symmetric models (Webster, 1983; Goswami and Shukla, 1984). The surface latent and sensible heat fluxes into the boundary layer that destabilizes the atmosphere ahead of the ascending zone were shown to play a key role in the generation and maintenance of the phenomenon (Webster, 1983). On the other hand, Goswami and Shukla (1984) stressed the role of convection–thermal relaxation feedback in northward propagation. Convective activity results in an increase of moist static stability so that convective activity is gradually self–destructed. Meanwhile the dynamical and radiative relaxation decrease moist static stability and bring the atmosphere to its original convectively unstable state. Anderson and Stevens (1987a) postulated a similar hypothesis.

2. *Energetics, Basic Dynamic Processes, and Diabatic Heating*

How are observed intraseasonal circulation anomalies propagating around the globe developed and maintained against dissipation? This question concerns the energetics of MJO. In spite of the difficulty in obtaining a reliable estimation of vertical motion over the tropical ocean, computations of energy conversion were attempted. In the Indian and western Pacific Oceans, the divergent wave associated with MJO is thermal direct with a net conversion of eddy available potential energy from diabatic heating over this frequent range (Murakami et al., 1984; Krishnamurti et al., 1985), albeit in the eastern Pacific the intraseasonal variability is possibly forced by high–frequency eddy kinetic energy conversion (Murakami and Nakazawa, 1985). On the basis of observed characteristic scales for MJO, it was shown that the intraseasonal variation in precipitation rate (about 3 mm day^{-1}) is necessary and sufficient to force an order of $O(5\times10^{-7}\mathrm{s}^{-1})$ divergence motion to maintain the vorticity balance required for maintaining the baroclinic structure of the Madden–Julian mode (Wang, 1988a). The above energy source arguments also provide an interpretation of the vertical structure. Since strong heating is released in the mid–troposphere, it follows that the oscillation must be of the gravest baroclinic mode structure, one of the fundamental observed features.

The basic dynamic processes are associated with fundamental modes in the tropical atmosphere. These include inertia–gravity waves, Rossby waves, Kelvin waves, and mixed Rossby–gravity waves as demonstrated by Matsuno's (1966) pioneering work. A unique feature in the tropics is the dynamic effect of the equator where the vertical component of the planetary vorticity changes its sign and its latitudinal variation maximizes. This results in feature distinct from those of midlatitude, i.e., the existence of equatorial Kelvin waves and mixed Rossby–gravity waves and the quantization of the inertia–gravity waves into a family of Poincare modes whose phase speeds are asymmetric w.r.t. the sense of phase propagation. All these free modes are admittedly involved in basic dynamic processes for tropical weather and low–frequency variations.

The observed Madden–Julian mode has a planetary zonal scale which is much larger than its meridional scale which is an order of equatorial Rossby radius of deformation. This anisotropic scale leads to a semi–geostrophic nature of the low–frequency motion, namely, the zonal wind is, to the first order, in geostrophic balance with pressure gradient forces. A direct consequence of this approximation (often referred to as thr long–wave approximation) is filtering out–high frequency inertia–gravity waves, mixed Rossby–gravity waves, and short Rossby waves. It simplifies the dynamic framework for the study of low–frequency motion. The essential dynamics of low frequency modes appear to be understandable without considering high frequency gravity modes.

What is the nature of the interrelationship between diabatic heating and basic dynamic processes? There have been two types of hypothesis. The first type regards MJO as a manifestation of the tropical atmospheric response to an imposed stationary heat source pulsating in intensity with a 40–50 day period (Yamagata and Hayashi, 1984; Anderson and Stephens, 1987b) or a moving heat source with either constant strength (Chao, 1987) or oscillating strength (Hayashi and Miyahara, 1987). The model of this type of hypothesis can be viewed as an extension of Gill's (1980) model by assuming the heat source pulsating in intensity or moving eastward. The weaknesses of this hypothesis are the failure in explaining the unstable development and the means of selecting the oscillation frequency and the failure in describing the feedback of atmospheric motion to heat source.

The second type of hypothesis considers heat sources interacting with large–scale flow. The idea originated from wave–CISK (Hayashi, 1970; Lindzen, 1974). Two types of diabatic heating representation have been proposed. One is evaporation heating which is assumed in linear dynamics to be proportional to the surface zonal–wind anomalies and was referred to as evaporation–wind feedback (Neelin et al., 1987; Emanuel, 1987). This mechanism was shown to be important to the MJO in an idealized GCM model, but the existence of the model oscillation does not depend on it (Neelin et al., 1987). The theoretical model assumed an ideal, neutrally stratified (gravity wave speed vanishes) atmosphere (which deliberately excludes the CISK mechanism) with pure zonal motion. In general, condensational heating can not be exactly canceled by adiabatic cooling. The model should admittedly have an intrinsic Kelvin mode, thus additional evaporational heating will enhance Kelvin wave–CISK. It is the east–west asymmetries in surface latent heat fluxes that play a key role, leading to eastward phase propagation and the growth of the unstable modes. The direction of the phase propagation depends on the direction of the mean flow. The assumption of a uniform easterly basic flow was questioned by Wang (1988b) but may be acceptable in regions of the central–eastern Pacific and Atlantic Oceans (Neelin, 1988).

Another type of heating representation follows conventional CISK theory, assuming that heating is controlled by low–level moisture convergence and may be referred to as condensation–convergence feedback. Both observation and numerical simulations appear to support the notion that the interaction between equatorial large–scale waves and organized moist convection plays an essential role in the generation and maintenance of these disturbances. The essence of the equatorial wave–CISK theory can be demonstrated in terms of prototype analytical models in which only the equatorial Kelvin wave is considered (Lau and Peng, 1987; Chang and Lim, 1988; Wang, 1988a) or both Kelvin and long Rossby waves coexist (Wang and Rui, 1990b).

3. *The Interpretation of the Madden–Julian Oscillation*

Any theoretical modeling of a complex geophysical phenomenon requires a focus on essential characteristics so that fundamental mechanisms can be identified. Among the many observed features of the Madden–Julian oscillation, the following may be viewed as essential and their mechanisms must be addressed by the theory:

(1) Unstable development: the disturbances (Madden–Julian waves) develop Y. Yover the Indian Ocean and western Pacific, and normally weaken as they pass the maritime continent and decay over the cold water east of the dateline. The frequency and intensity of eastward–moving disturbances are greater in boreal winter than in boreal summer.

(2) Movement: the disturbances move eastward slowly in the Eastern Hemisphere (when convection and circulation anomalies couple) at a speed of about 5 m s^{-1} and about three times faster in the Western Hemisphere after circulation anomalies decouple from convection.

(3) Vertical structure: the zonal wind fluctuations in the upper troposphere are nearly out of phase with and much stronger (by a factor of two) than those in the lower troposphere. The convection anomalies are nearly in phase with the equatorial low–level westerly anomalies.

(4) Horizontal structure: the circulation is characterized by dominant zonal winds near the equator and substantial meridional winds off the equator. The equatorial upper–(lower–)level easterlies (westerlies) are coupled with twin subtropical anticyclonic (cyclonic) circulation cells.

(5) Horizontal scale: the circulation has a planetary zonal scale with an ascending area significantly smaller and stronger than adjacent descending branches.

In spite of quantitative discrepancies, many of the above–mentioned basic features are qualitatively explainable in terms of simple or intermediate wave–CISK models.

The preferred development over the warm water of the Indian and western Pacific Oceans can be explained by the instability of the moist tropical atmosphere to large–scale perturbations. This instability which depends on the vertical distribution of the moist static energy of basic flow, requires that the underlying sea surface temperature (SST) exceeds a critical value (Wang, 1988a). Failure to satisfy this requirement leads to decay of the MJO over the cold water east of the dateline. The zonal variation of SST along the equator modulates the convective activity and the development, creating the contrast in the strength of the intraseasonal variability between the Eastern and Western Hemispheres (Miyahara, 1987; Lau and Peng, 1987). Why the disturbances tend to be dissipated or to split near the maritime continent (which is referred to as convective energy source) remains unclear. The model results also demonstrate that when maximum SST moves from the equator to 7.5°N, the growth rate of the unstable waves is significantly reduced, suggesting that the annual march of the "thermal equator" and associated convective heating is likely responsible for the annual variations of the equatorial 40–50 day wave activities (Wang and Rui, 1990b).

The oscillation period is closely associated with the phase speed of the Madden–Julian waves. The slow eastward propagation may be caused by a number of mechanisms: the inclusion of viscosity (Chang, 1977), the interaction between internal vertical wave–CISK modes (Chang and Lim, 1988; Wang and Chen, 1989), the reduction of the static stability due to latent heat release (Swinbank et al., 1988; N. C. Lau et al., 1988), the heating maximized in the lower troposphere between 500 and 700 hPa (Takahashi, 1987; Sui and Lau, 1989), and the coupling between moist Kelvin and Rossby modes (Wang and Rui, 1990b). Without or with little convective heating the disturbances would move much faster such as observed in the Western Hemisphere. The dominant eastward propagation is likely associated with suppression of Rossby–wave response to the west and enhanced Kelvin–wave response to the east of the propagating heat source due to positive–only heating (Lau and Peng, 1987), or due to the nonlinearity which leads to rapid stabilization of the atmosphere to the west of the CISK heating (Hendon, 1988).

The out–of–phase relationship between the upper and lower tropospheric zonal winds arises from the maximum latent–heat release in the middle troposphere. The asymmetry in vertical structure with the largest amplitude occurring in the upper troposphere can be caused by vertical model coupling through convective heating associated with the boundary–layer moisture convergence (Wang and Chen, 1989). This is confirmed by the fact that the intraseasonal disturbances simulated by a general circulation model with boundary–layer friction exhibit asymmetric vertical structures (N. C. Lau et al., 1988), whereas inviscid multi–level wave–CISK models without a boundary layer do not produce this asymmetry (e. g., Lau and Peng, 1987). None of the models to date explains why the observed convection anomalies are nearly in phase with low–level westerly anomalies.

The dominance of the zonal wind components near the equator and the existence of significant meridional wind components off the equator suggest that the observed horizontal structure resembles a coupled Kelvin–Rossby mode. In fact, the presence of heating does not allow a pure zonal motion since heat–induced overturning always generates significant off–equator meridional flow. An idealized moist Kelvin wave model is merely instrumental in illustrating the wave–CISK mechanism and may be viewed as a first–order approximate solution valid at the equatorial zonal plane. The possible processes responsible for the coupling between the Kelvin and Rossby waves include nonlinear advection (Hendon, 1988) and convective heating induced by boundary–layer moisture convergence (Wang and Rui, 1990b). The present inability to model the rotational wind component with realistic strength may suggest the importance of the inhomogeneity of the mean state on the wave motion which has not been adequately included in the theoretical modeling.

The linear wave–CISK theory failed to account for the selection of planetary–scale motion because of explosive growth of shortwave components. Lim et al. (1990) argued that CISK mechanism inherently possesses a severe form of nonlinearity that takes full effect for even infinitesimal perturbations. The positive–only heating is able to produce the exponentially growing wavenumber–one flow patterns that propagate without change of shape (Lau and Peng, 1987). Itoh's (1989) numerical experiments confirmed the effect of positive–only heating. In order for a super cluster to dominate among many possibly growing cloud clusters, dry regions were assumed to exist over wide areas in which the generation of cumulus by weak moisture convergence was artificially prohibited. Using a single vertical mode Kelvin wave–CISK model, Wang and Xue (1991) have shown that the unstable wave packet with nonlinear heating is characterized by two zonal–length scales: the ascending branch length (ABL) and total circulation extent (TCE). For a given basic state, the growth rate of a wave packet increases with decreasing ABL or TCE. However, up to a moderate growth rate (order of day^{-1}) the energy spectra of all wave packets are dominated by zonal wavenumber–one regardless of ABL size. In particular, the slowly growing (low frequency) wave packets normally exhibit TCEs of planetary scale and ABLs of synoptic scale. This may account for why observed equatorial intraseasonal disturbances often display a narrow convection region in between two much broader dry regions and a total circulation of planetary scale.

4. *Other Dynamic Aspects of the Madden–Julian Oscillation*

The instability theory does not address the question of what triggers the initial formation over equatorial Africa and rapid development over the Indian Ocean of the low–frequency disturbances. Three processes, however, may possibly act as triggers. First, the upper–level divergence waves that travel across the Western Hemisphere after decoupling with convection (Knutson and Weickmann, 1987) may induce subsequent convection burst over the equatorial Africa or Indian Ocean. Another possible propagation route starts from the central equatorial Pacific emanating away from the equator across South America and the Atlantic turning equatorward to the equatorial Indian Ocean (Gao and Stanford, 1988). The last possible trigger is due to active interaction between extratropics and tropics. Strong equatorward flows in the western Indian Ocean and western Pacific may activate equatorial convection in those regions (Murakami, 1987, 1988). Hsu et al. (1990) and Li and Xiao (1990) suggested that some intraseasonal oscillation events may be initiated by Rossby wave propagation from higher latitude into the equatorial regions.

Wang and Rui (1990a) pointed out that about one–half of the northward propagation over the Indian Ocean and western Pacific regions are not associated with the equatorial eastward moving systems. The existence of independent northward–moving disturbances suggests that the mechanism responsible for meridional propagation may differ from that of eastward propagation. For the northward propagation, the interaction of monsoonal large–scale circulation with intraseasonal modes appears to play an important role. Lau and Peng (1990) has shown that as a result of this interaction unstable baroclinic disturbances with periods of 5–6 days are generated over the monsoon region along 15–20°N. The development of monsoon disturbances and simultaneous weakening of the equatorial Madden–Julian mode are accompanied by a rapid northward shift of the rising branch of the local Hadley cell. The northward propagation was also interpreted by Chen and Liou (1991) using a quasi–balanced model in which the low frequency motion is assumed to be in diabatic balance. The balanced wind–field forced by monsoon heating has a remarkable northward component in the lower branch. This balanced wind advection in turn forces the quasi–balanced wave to move northward.

The Madden–Julian oscillation is a multi–scale process. Lau et al. (1989) showed that in the formative stage of the 30–60 day disturbances (which can be identified as a group of super cloud clusters) westward propagating cloud cells exist which are of Rossby waves in nature (also see Yamagata, 1987). When the atmospheric wave is slowed remarkably to a few meters per second by condensation–moisture convergence feedback, the atmospheric moist Kelvin–wave may be destabilized by air–sea coupling (Lau and Shen, 1988; Hirst and Lau, 1990), i.e., the atmosphere–ocean interaction may amplify intraseasonal disturbances of internal atmospheric origin. In such a case direct wind–evaporation feedback would be more important than the coupled Kelvin–wave instability.

V. SUMMARY AND PERSPECTIVE

Our knowledge of intraseasonal variability of the tropical atmosphere has increased tremendously in the last decade. An observational basis has been firmly established for various aspects of the Madden–Julian oscillation (MJO). Tropical intraseasonal variations are viewed as far more complex today than a decade ago. Our understanding of the dynamics of MJO is in continual progress, yet remains rather limited.

The planetary–scale Madden–Julian oscillation modulates and interacts with higher frequency disturbances. For instance, the Madden–Julian waves appear to be phase–locked with 10–20 day medium–scale disturbances that travel westward (Krishnamurti et al., 1985).

The contribution of organized storms and cyclones to the intraseasonal variability in the western Pacific–Indian Ocean region is substantial (Smith and Mehta, 1990). The interactions among planetary–scale MJO and synoptical–scale long–lasting super cloud clusters, storms and cyclones, and high–frequency cloud clusters deserve further investigation.

Tropical mean flows have remarkable season–dependent zonal, meridional, and vertical variations. Understanding of the interactions of the Madden–Julian oscillation with the mean flows is of central importance for the monsoon dynamics and teleconnection. The MJO is found to have remarkable influences on the onset, break, and retreat of the Indian monsoon (Krishnamurti and Subrahmanyam, 1982; Murakami et al., 1986; Chen et al., 1988) as well as the Australian monsoon (Hendon and Liebmann, 1990). The relationship between the regional westward propagation in the Indian and western Pacific monsoon domain and global eastward propagation on an intraseasonal time–scale needs further studies. The global influences of the MJO are also noteworthy. The interaction and association between the MJO and variations in midlatitude circulation are evident (Weickmann, 1983; Liebmann and Hartmann, 1984; Lau and Phillips, 1986; Ghil and Mo, 1991a,b) but the nature of this interaction remains controversial.

Empirical evidence appears to suggest that the air–sea interaction may be acting on an intraseasonal time–scale (Krishnamurti et al., 1988). The oceanic influences on MJO remain unraveled. The existence of significant intraseasonal variations in the eastern equatorial Pacific and along the west coast of America has been documented and postulated to be remotely forced by the counterpart oscillation in the atmosphere over the western Pacific (e. g., Spillane et al., 1987; Enfield, 1987). Similar oscillation signals were found in the equatorial Indian Ocean (e.g., Mertz and Mysak, 1984). Unfortunately, lack of oceanic observations over the western Pacific and Indian Oceans has hampered further investigation of the nature of air–sea interactions on this time scale. The on–going TOGA COARE project is expected to promote the research in this regard.

The possible roles Madden–Julian mode may play in triggering Pacific warm events and some causal relationships arising from the interaction among annual cycle, MJO and ENSO were speculated based on empirical analyses and simple coupled model analysis (Nitta and Motoki, 1987; Lau and Chan, 1988; Lau and Shen, 1988). Since the initiation and termination of tropical Pacific warm events occur on a relatively short time scale of a few months, and since the rapid changes in atmospheric conditions are very likely the cause of irregularity of ENSO cycle (Philander, 1990), study of the links between two major low frequency variabilities is much needed for prediction of ENSO.

In view of the fact that the intraseasonal variations are linked to a variety of fundamental physical processes, a thorough understanding of the phenomenon has potential impact on prediction of long–range variations of the tropical atmospheric circulation. Such experiments have been undertaken recently (Krishnamurti et al., 1990). Prediction of the Madden–Julian oscillation is a rather challenging task and will be one of the major breakthroughs in numerical modeling of the general atmospheric circulation and climate dynamics.

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