Structures of the MJO: Implications to Its Air-Sea Interaction and Dynamics

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Abstract

Based on observations, theories, and numerical simulations, the Madden-Julian Oscillation (MJO) may take different structures in terms of the phase between its large-scale convective center and surface winds. This study explores the implications of the MJO structure to its dynamics and air-sea interaction, especially feedback from sea surface temperature (SST). A scale analysis demonstrated that if SST feedback is important to the MJO, the MJO must be highly sensitive to latent heat flux. This sensitivity can be further amplified by the differences in the MJO structures. The phase between deep convection and surface winds of the MJO determines the phase between the surface winds and SST, which in turn results in either enhancement or reduction of the intraseasonal fluctuations in the latent heat flux. The dynamics that might be responsible for different MJO structures are then discussed in terms of interaction between deep convection and the large-scale circulation, and scale interaction between the Kelvin-Rossby wave complex of the MJO and mesoscale momentum transport. It is concluded that understanding and simulating correctly the MJO structure is pivotal to the study of air-sea interaction of the MJO. It is recommended that the scale interaction of the MJO and its sensitivity to latent heat flux receive more research attention.

1. Introduction

Recent studies (Zhang and McPhaden 2000; Zhang and Anderson 2003) have summarized the structure of the Madden-Julian Oscillation (MJO) (Madden and Julian 1971, 1972) into four idealized models. Each model represents a particular phase between the convective center and surface zonal wind of the MJO (Fig. 1). Models I and II are commonly observed in the equatorial Indian and western Pacific Oceans (e.g., Rui and Wang 1990; Inness and Slingo 2003; Sperber 2003). Model III is predicted by an MJO theory (Emanuel 1987; Neelin et al. 1987). Model IV is commonly simulated numerically (e.g., Wang and Li 1994; Maloney and Hartmann 2001). Neither Model III nor IV has been found from observations. The amplitude of intraseasonal fluctuations in sea surface temperature (SST) induced by the MJO can vary because of these different structures (Zhang and Anderson 1993). The intraseasonal fluctuations in SST are mainly determined by those in surface solar radiation and latent heat fluxes (Anderson et al. 1996; Shinoda and Hendon 1998; 2001). Latent heat flux is mainly controlled by surface winds (Hartmann and Michelson 1993). In MJO Model II, for example, surface cooling by the reduction in solar radiation flux due to cloudiness in the convective center coincides with cooling by enhanced latent heat flux due to the maximum zonal wind. This leads to a large SST perturbation. In contrast, cooling of solar radiation is nearly balanced by reduced latent heat flux in Model IV, resulting in a negligibly small SST perturbation. The phase between the SST perturbation and surface zonal wind also depends on the MJO structure. This may affect possible feedback from the ocean to the atmosphere. Such feedback has been suggested by numerical simulations of the MJO with an interactive ocean (e.g., Flatau et al. 1997; Waliser et al. 1999; Inness and Slingo 2003). The mechanisms for such feedback remain unknown. The first objective of this study is to explore how the phase differences due to the structural changes of the MJO may affect possible SST feedbacks.

The different structures of the MJO as depicted by the four idealized models in Fig. 1 also pose a challenge to the understanding of the MJO dynamics. Most efforts of understanding the MJO have been focused on explaining the frequency selection and slow eastward propagation of the MJO. The dynamics that determines the structure of the MJO, especially the phase between its convective and wind component, needs to be explored, which is the second objective of this study.

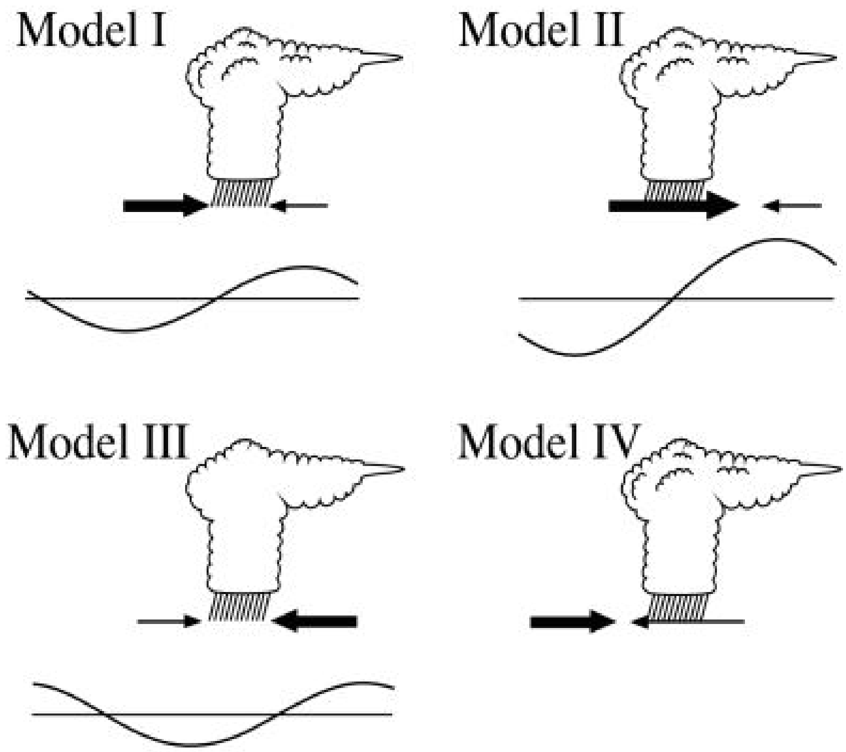

Figure 1. Sketches of idealized models for surface structures of the MJO. The precipitating cloud symbols represent the large-scale centers of deep convection and precipitation of the MJO, which is also the location of minimum solar radiation fluxes at the surface. The horizontal arrows represent surface zonal winds associated with the MJO; right-pointing ones denoting westerlies and left-pointing ones easterlies; the location of maximum surface wind speed is indicated by the thick arrows, where is also the location of maximum surface fluxes of latent and sensible heat from the ocean. SST perturbations associated with each model are illustrated at the bottom of each panel except for Model IV (see Section 3 for details). The exact phase lag between maximum surface zonal wind (maximum surface fluxes) and maximum precipitation (minimum solar radiation) is 90° in Model I, 0° in Model II, -90° in Model III, and 150° in Model IV. (From Zhang and Anderson 2003)

In section 2, a scale analysis based on observations from the Tropical Ocean-Atmosphere (TAO) mooring array (McPhaden et al. 1998) first demonstrates that the feedback from the intraseasonal perturbations in SST to the MJO is unlikely related to the nonlinear Clausius-Clapeyron effect. The amplitude of the intraseasonal perturbations in SST is too small for the nonlinearity to be affective. The scale analysis continues to show that fluctuations in latent heat flux due solely to the intraseasonal perturbations in SST are also very small in comparison to the total intraseasonal perturbations. This implies that the MJO must be very sensitive to perturbations in latent heat flux, if the ocean indeed feeds back to the MJO.

A simple analytical approach in section 3 illustrates that the modification on the amplitude and phase of intraseasonal perturbations in latent heat flux by the intraseasonal perturbations in SST depends on the phase between SST and surface winds. Applying this to the four idealized MJO models, discussions are given on how SST feedback may increase or decrease the amplitude of intraseasonal perturbations in latent heat flux.

In section 4, possible dynamical reasons for each of the four idealized MJO structures are conceptually explored. Roles of moisture convergence, mesoscale momentum transport, and large-scale equatorial waves are discussed. Finally, concluding remarks are given in section 5.

2. Scale analysis

Any SST feedback to the atmosphere must go through changing surface energy fluxes. On the intraseasonal timescales, perturbations in latent heat flux are much larger than in sensible heat flux (Zhang 1996; Zhang and McPhaden 2000). In numerical models and observational analyses, latent and sensible heat fluxes (Q_l and Q_s) are commonly estimated using in bulk aerodynamic formula:

$$Q_l = \rho C_e L V \Delta q \tag{1}$$

$$Q_s = \rho C_h C_p V \Delta T \tag{2}$$

where C_e and C_h are the transfer coefficients, ρ is the air density, L the latent heat of evaporation, C_p the specific heat of moist air, V the wind speed, $\Delta q = q_s - q_a$ the air-sea humidity difference, and $\Delta T = T - T_0$ the air-sea temperature difference, with q_s and T being surface saturation humidity and temperature ($q_s = q_s$(T)) and q and T_a air humidity and temperature near the surface (e.g., at 10 m).

2.1. Sensitivity of the saturation vapor pressure

The importance of SST in the western Pacific warm pool is often argued in terms of the nonlinear dependence of the saturation vapor pressure on temperature as expressed by the Clausius-Clapeyron equation (Webster 1994)

$$e_s(T) = e_s(T_o)\, e^{\left[\frac{L}{R_v}\left(\frac{1}{T_o}-\frac{1}{T}\right)\right]} \tag{3}$$

where $e_s(T_0)$ is the saturation vapor pressure at a reference temperature T_0, L is the latent heat of evaporation or sublimation, and R_v is the specific gas constant for water vapor. Because of the nonlinearity in (3), the same change in T would result in a larger change in e_s at higher T than at lower T. The rate of change in the saturation vapor pressure with temperature is

$$\frac{de_s}{dT} = e_s(T)\frac{L}{R_v T^2} \tag{4}$$

With $L = 2.5 \times 10^6$ J kg^{-1}, $R_v = 461$ J K^{-1} kg^{-1}, and $T_0 = 273.5$ K, we have $de_s/dT = 2.45$ hPa K^{-1} at $T = 302.5$ K. For air in touch with the sea surface, T in (3) and (4) represents SST; for air near but not in touch with the sea surface, T is the air temperature. In both cases, $dT \leq 0.5$ °C in association with the MJO. The corresponding magnitude of the fluctuations in the saturation vapor pressure is thus 1.22 hPa, 3% of its observed mean (38 hPa) in the equatorial western Pacific and less than its observed one standard deviation (1.5 hPa) under all conditions there. The sensitivity is weak simply because the amplitude of the intraseasonal variations in SST is not large enough for the nonlinearity in (3) to be effective.

2.2. Sensitivity of latent heat flux and evaporation

The sensitivity of latent heat flux to SST is

$$\frac{\partial Q_l}{\partial T} = Q_l \frac{L}{R_v T^2} \tag{5}$$

for fixed relative humidity and exchange coefficient (Hartmann and Michelson 1993). This sensitivity actually stems from that of the surface saturation vapor pressure (4). With the mean values Q_l = 110 W m^{-2}, T = 302.5 K in the equatorial western Pacific, this sensitivity is 6.5 W m^{-2} K^{-1}. This is similar to that estimated by Hartmann and Michelson (1993) who used slightly different values of Q_l and T. The magnitude of fluctuations in latent heat flux solely due to the fluctuation in SST in association with the MJO is therefore 3.3 W m^{-2}. It is 3% of the mean (110 W m^{-2}) and 8 % of its total fluctuation on the intraseasonal timescales (~40 W m^{-2}).

The sensitivity of evaporation $E = Q_l/\rho L$ to SST can be similarly estimated. It is 0.22 mm day^{-1} K^{-1} at mean SST of 302.5 K, in comparison to mean E of 3.7 mm day^{-1} (equivalent to mean Q_l of 110 W m^{-2}). The magnitude of fluctuations in evaporation solely due to the fluctuation in SST in association with the MJO is therefore 0.11 mm day^{-1}, 3% of the observed amplitude of its intraseasonal variation. Assume water vapor evaporated into the atmosphere is uniformly distributed in a surface layer of 3-m deep (the height of TAO humidity sensors). Corresponding to the fluctuation in evaporation of 0.11 mm day^{-1} over an MJO cycle, the magnitude of changes in specific humidity would be 0.05 g kg^{-1} and the equivalent magnitude of changes in relative humidity would be about 0.7%. They are about 13% and 47% of the respective intraseasonal standard deviations observed from the TAO array measurement (Zhang 1996). Apparently, less than one-third of the intraseasonal variability of the surface relative humidity can be related to the fluctuation in moisture, of which only slightly more than 10% of the variability can directly be related to the intraseasonal variations in SST.

Because latent heat flux also varies with surface wind speed and humidity, its sensitivity to SST cannot be fully perceived without being compared to its sensitivities to these two variables. The overall sensitivity of latent heat flux is

$$\begin{aligned} dQ_l &= \frac{\partial Q_l}{\partial V} dV + \frac{\partial Q_l}{\partial T} dT + \frac{\partial Q_l}{\partial q} dq \\ &= \frac{Q_l}{V} dV + \frac{Q_l L}{RT^2} dT + \frac{Q_l}{\Delta q} dq \end{aligned} \tag{6}$$

where $\frac{\partial Q_l}{\partial V} = \frac{Q_l}{V}, \frac{\partial Q_l}{\partial T} = \frac{Q_l L}{RT^2}, \text{and} \frac{\partial Q_l}{\partial q} = -\frac{Q_l}{\partial q}$ are the sensitivities of latent heat flux to wind speed, SST, and surface air specific humidity, respectively, at fixed density and exchange coefficient. Using the mean values of these variables from the TAO observations in the equatorial western Pacific, one can estimate those sensitivities as

$$\frac{\partial Q_l}{\partial V} = \frac{Q_l}{V} \approx 26 \mathrm{Wm^{-2} (ms^{-1})^{-1}} \tag{7a}$$

$$\frac{\partial Q_l}{\partial T} = \frac{Q_l L}{RT^2} \approx 7 \mathrm{Wm^{-2} K^{-1}} \tag{7b}$$

$$\frac{\partial Q_l}{\partial q} = \frac{Q_l}{\Delta q} \approx -18 \mathrm{Wm^{-2} (g\ kg^{-1})^{-1}} \tag{7c}$$

Thus, the sensitivity is the strongest to the wind speed and the weakest to SST. The change of 7 W m^{-2} in latent heat flux caused by a maximum change of 1°C in SST associated with the MJO can be equally induced by a change of 0.3 m s^{-1} in wind speed and a change of 0.4 g kg^{-1} in surface air specific humidity. While 1 °C is much larger than the intraseasonal standard deviation of SST, 0.3 m s^{-1} (0.4 g kg^{-1}) is much smaller than the intraseasonal standard deviation of wind speed (surface air specific humidity). If these standard deviations estimated from the TAO observations for the western Pacific were independent of each other, then the fluctuations in latent heat flux caused by the variability of SST would be about 10% and 20% of those caused respectively by the variability of wind speed and surface air humidity.

The above scale analysis suggests that effects of the intraseasonal fluctuations in SST on latent heat flux (also sensible heat flux) appear to be weak for two reasons. First, the magnitude of the intraseasonal variations in SST is not large enough for the nonlinearity in the Clausius-Clapeyron equation to be effective. Second, the variability of surface latent and sensible heat fluxes is controlled more by surface winds and by SST. The analysis, however, does not take into account a secondary effect of the variability in surface wind speed V, namely, its phase relationship with SST. From the bulk algorithm (1), $\frac{\partial Q_l}{\partial T} = \frac{Q_l L}{RT^2}$ is proportional to V. A varying amplitude of V (0 - 5 m s^{-1}) during a cycle of an MJO event would inevitably modulate the sensitivity of Q_l to SST. This modulation due to the relative phases between SST and V is illustrated in the next section.

3. MJO structure and SST feedback

Surface wind speed V and air-sea humidity difference Δq in (1) can be decomposed into their intraseasonal components, V_i and Δq_i, and non-intraseasonal components representing the rest of their variability, V^* and Δq^*:

$$V = V^* + V_i \tag{8a}$$

and

$$\Delta q = \Delta q^* + \Delta q_i \tag{8b}$$

Assume the intraseasonal components are of simple forms,

$$V_i = A\sin(t) \tag{9a}$$

and

$$\Delta q_i = B\sin(t+\alpha) \tag{9b}$$

where A and B are amplitudes of the intraseasonal fluctuations in V and Δq, respectively, t is time, and α is the phase lag between V_i and Δq_i. The intraseasonal variability of latent heat flux (Q_i) can then be expressed as

$$Q_i = a\gamma\sin(t+\varphi) \tag{10}$$

where

$$\gamma = \left[1 + 2c\cos(\alpha) + c^2\right]^{1/2} \tag{11}$$

is an amplitude modification factor of Q_i due to Δq_i ($\gamma = 1$ if $\Delta q_i = 0$) and

$$\varphi = \tan^{-1}\left[\frac{c\sin(\alpha)}{c\cos(\alpha)+1}\right] \tag{12}$$

is the phase lag between Q_i and V_i because of Δq_i ($\varphi = 0$ if $\Delta q_i = 0$). In (11) and (12*), $c = BV^*\Delta q^*/A$* is the ratio between the amplitudes of fluctuations in Q_i solely due to V_i and Δq_i, respectively. If intraseasonal perturbations in the air humidity are neglected, which is not a bad assumption (Zhang 1996), then Δq_i is solely due to the intraseasonal perturbations in SST (SST_i) and the effects of SST_i on latent heat flux are measured by γ and φ.

In Fig. 2, γ and φ are plotted as functions of α and c. The amplitude of Q_i may be either amplified ($\gamma > 0$) or reduced ($\gamma < 0$) by SST_i, depending on both α and c. Take MJO Model I (Fig. 1) as an example. In this model, SST_i is almost completely out of phase with V_i ($\alpha \leq 180°$) and it reduces the amplitude of intraseasonal fluctuations in latent heat flux ($\gamma < 0$). This is consistent to the result from Shinoda et al. (1998) based on a global model reanalysis product. In contrast, for MJO Model III, SST_i is almost completely in phase with V_i ($\alpha \approx 0°$) and it amplifies the amplitude of intraseasonal fluctuations in latent heat flux ($\gamma > 0$). For Model II, SST_i leads V_i by about a quarter of cycle ($\alpha \approx -90°$) and its effect on the amplitude of intraseasonal fluctuations in latent heat flux is very weak ($\gamma \approx 0$).

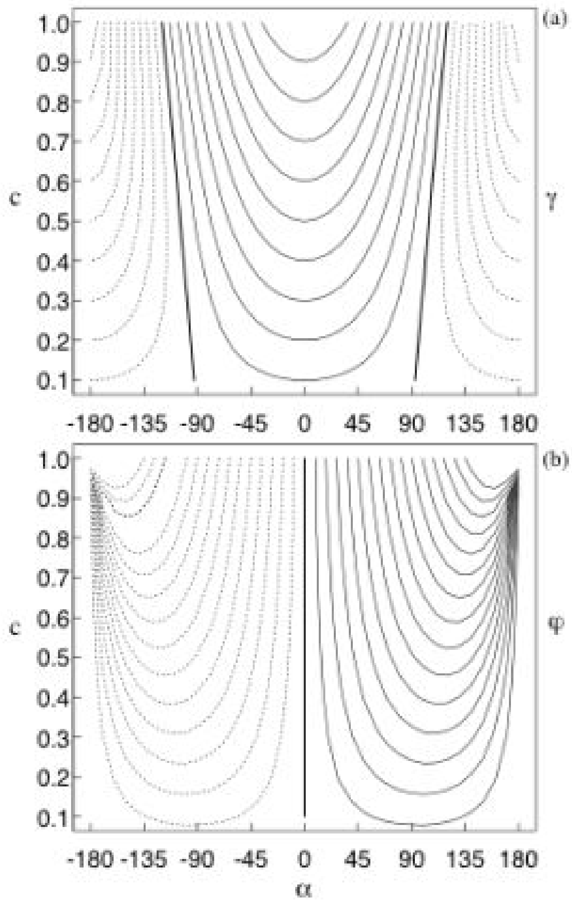

Figure 2. (a) Amplitude modification factor (g) and (b) phase lag between Qi and Vi (j) due to Δqi (see definitions in Equ. 10-12) as functions of the amplitude ratio, and (c) phase difference between Vi and Δqi (a). Negative values are plotted by dotted lines and zeros by heavy solid lines. Contour intervals are 0.1 in (a) and p/40 in (b).

4. MJO structure and dynamics

If the SST feedback to the MJO sensitively depends on the MJO structure, than it would be extremely important to simulate correctly the MJO structure in numerical and theoretical models used as tools to study air-sea interaction of the MJO. Equally important is to understand the dynamics that may lead to a particular structure of the MJO. Among the four idealized MJO structural models summarized in Fig. 1, some are better understood than others. Model I, which represents the classic structure as described by Madden and Julian (1972), can be understood in terms of either deep convection located in the region of the strongest surface moisture convergence due to the equatorial Kelvin wave which is the dynamic backbone of the MJO (Wang 1988) or the large-scale circulation (in terms of the Kelvin and Rossby waves) responding to deep convective heating (Webster 1972; Gill 1980), as illustrated in Fig. 3a. Imposing Model I on to an easterly mean flow would transform it to Model III. Boundary-layer friction would shift surface moisture convergence eastward into the surface easterly region. If a model atmosphere is over sensitive to surface moisture convergence, this may give rise to Model IV.

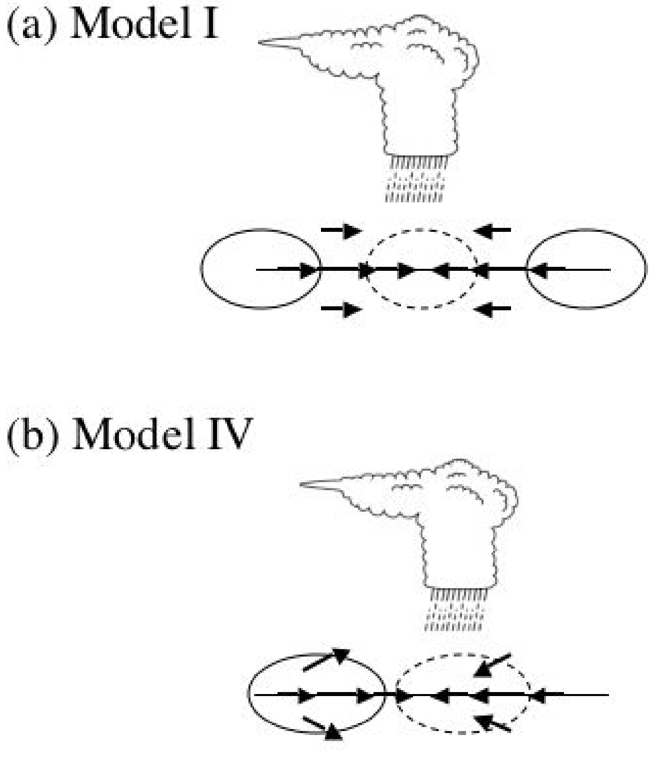

Figure 3 Schematic diagram illustrating possible dynamics for the structure of (a) idealized MJO Model I and (b) Model IV. Arrows indicate surface zonal wind component of the equatorial Kelvin wave, contours indicate surface wind divergence with dashed contours being negative (convergence). Symbols of precipitating cloud mark the location of the large-scale convective center relative to the surface zonal wind.

Model II appears to be difficult to explain purely in terms of the large-scale dynamics. A steady-state response of the atmosphere to deep convective heating indeed places low-level westerlies through two-third of the convective region (e.g., Gill 1980). But this is insufficient to explain the observed dominance of surface and low-level westerlies in the convective region of the MJO (Zhang 1996; Zhang and McPhaden 2000; Inness and Slingo 2003; Sperber 2003). It is particularly puzzling to realize that surface westerlies promote divergence near the equator. Scale interaction between the large-scale circulation and mesoscale convective systems may play pivotal roles in determining the structure of MJO Model II.

First, it has been well observed that embedded within the eastward moving large-scale convective envelope coupled with the large-scale circulation of the MJO are westward moving synoptic and mesoscale convective systems (e.g., Nakazawa 1988; Chen et al. 1996). It is possible that these westward moving convective systems help shift the large-scale envelop, also known as super cloud clusters, westward to overlap with surface westerlies (Fig. 4). A more convincing argument explaining the collocation of large-scale convective envelope and surface westerlies has been put forth by Houze et al (2001). In this argument, mesoscale convective systems develop into a mature stage with dominant stratiform precipitation and a mid-level (below the freezing level near 5.5 km) rear inflow. If deep convection starts as in Model I (Fig. 5a), the mid-level rear inflow would be enhanced by the mean large-scale westerlies that is part of the Kelvin-Rossby wave complex of the MJO (Wang and Rui 1990). Meanwhile, stratiform precipitation, mainly ice particles and supper cool raindrops, falls into the rear inflow and makes it descend by diabatic cooling of melting, sublimation, and evaporation. The descending rear inflow enhances surface westerly (Fig. 5b). By this mesoscale downward momentum transport, surface westerlies are collocated with the strongest deep convection associated with the MJO. The key to this process is the Kelvin-Rossby wave complex that provides mean low and mid-level westerlies west of the large-scale convective envelope.

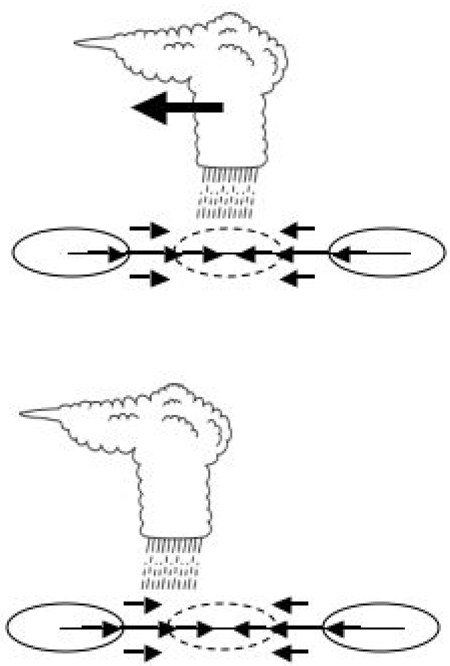

Figure 4 Same as Fig. 3 except for MJO Model II with the westward propagation of synoptic and mesoscale convection systems marked by the thick arrow.

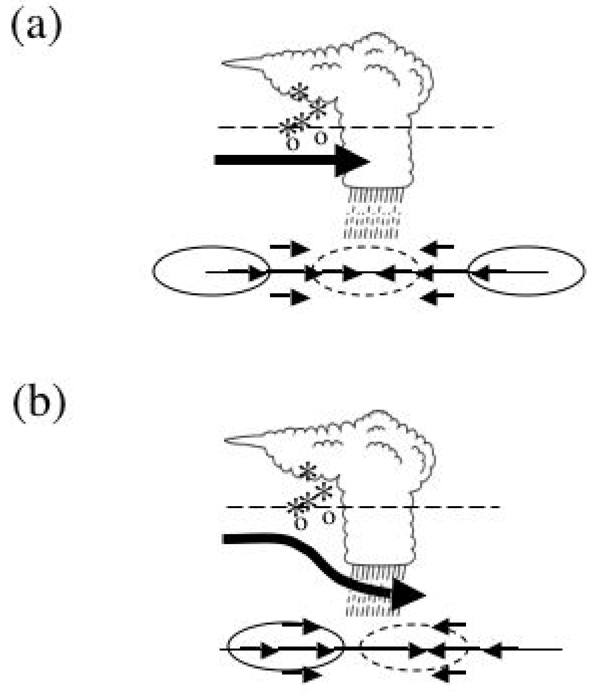

Figure 5 Same as Fig. 3 except for MJO Model II with stratiform precipitation represented by circles and asterisks, the rear inflow by the thick arrows, and the freezing level by the dashed lines. (see Houze et al. 2001)

5. Concluding remarks

This study has shown that if SST feedback to the MJO is important, then the MJO must be highly sensitive to small changes in surface latent heat flux. This sensitivity comes from two factors: the small amplitude of the intraseasonal fluctuation in SST, and the dependence of the SST modulation to latent heat flux on the structure of the MJO. It is also shown that the dynamics responsible for the MJO structure must be understood in terms of not only interaction between the large-scale circulation and deep convection but also scale interaction between the Kelvin-Rossby wave complex and mesoscale convective systems. The mesoscale momentum transport was shown to be an excellent example of such scale interaction. The sensitivity to latent heat flux and the scale interaction of the MJO are two subjects that deserve more research attention.

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