

Cross-Scale Air-Sea Interaction in the Tropics Related to the MJO

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1. Introduction

This article discusses two examples of cross-scale air-sea interaction in the tropics involving the Madden-Julian Oscillation (MJO, Madden and Julian 1972). Figure 1 sketches these two examples. In addition to inducing intraseasonal perturbations in sea surface temperature (SST), the MJO also interacts with the diurnal cycle in SST and atmospheric moist convection, as illustrated in the left part of Fig. 1, and with El Niño – Southern Oscillation (ENSO), illustrated in the right part of Fig. 1. Detailed processes in these two examples are discussed in sections 2 and 3, respectively. Conclusion remarks are given in section 4. For other discussions on the subject of air-sea interaction involving the MJO and phenomena of other scales, see reviews by Hendon, Kessler, and Lau in Lau and Waliser (2005) and by Zhang (2005).

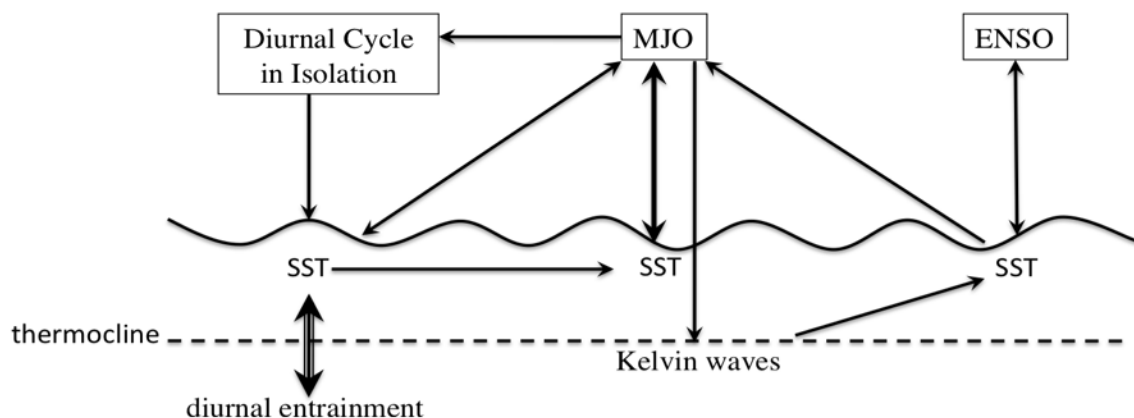


Figure 1: Schematic diagram of cross-scale air-sea interactions between the MJO and diurnal cycle and between the MJO and ENSO. Arrows denote directions of influences.

2. MJO and Diurnal Cycle

Systematic studies on interaction between the MJO and diurnal cycle mainly started from TOGA COARE¹. Recent observations and numerical modeling have substantially improved our understanding of this issue. Discussions in this subsection constitute a brief review of our current knowledge largely based on the following studies, unless specially indicated otherwise: Anderson et al (1996), Weller and Anderson (1996), Cronin and MaPhaden (1997), Shinoda and Hendon (1998), Singlo et al (2003), Zhang and Anderson (2003), Bernie et al. (2005, 2008).

In the equatorial western Pacific warm pool region, where the MJO is strong, fluctuations in mixed-layer temperature (T) on diurnal and intraseasonal timescales are roughly governed by one-dimensional (1-D)

¹ Tropical Ocean Global Atmosphere Coupled Ocean-Atmosphere Research Experiment, November 1992 – February 1993 (Webster and Lukas 1993)

processes that can be approximately described by the following equation

$$\frac{\partial T}{\partial t} = \frac{1}{\rho C_p h} (Q - Q_{sh}) - \frac{T - T_h}{h} \left(\frac{dh}{dt} + w_h \right) \quad (1)$$

where t is time, ρ water density, C_p water heat capacity at constant pressure, h the depth of the mixed layer, Q net surface heat flux, Q_{sh} solar radiation penetrating through the base of the mixed layer, T_h temperature at the base of the mixed layer, and w_h the vertical velocity at the base of the mixed layer. The bulk diurnal and intraseasonal variations in SST are also governed by this equation with exceptions that will be discussed later in this subsection.

The first systematic observation of an intraseasonal modulation of the diurnal cycle in air-sea fluxes of energy was made using data from a moored buoy in the western Pacific warm pool. It clearly revealed the contrast in the diurnal cycle of the air-sea fluxes between periods of low and high winds that correspond to known phases of the MJO. During its convectively inactive phases, the diurnal cycle in insolation (day time solar heating) is the strongest because of anomalously low cloud cover. Meanwhile, because of weak mechanic mixing in the upper ocean due to anomalously low surface wind, the mixed layer is very shallow ($h < 20$ m). This makes both surface heating/cooling (the first term on the right hand side of Equ. 1) and entrainment through the base of the mixed layer (the second term) very effective in inducing changes in SST. The shallow mixed layer also allows entrainment cooling be easily provoked by convective mixing due to night time surface longwave radiation. In consequence, the diurnal cycle in SST is most robust while its intraseasonal tendency is positive. The opposite occurs during convectively active phases of the MJO. Weakened daytime solar heating due to anomalously high cloud cover and deepened mixed layer due to strong surface wind help suppress the diurnal cycle in SST when its intraseasonal tendency is negative.

This apparently robust heating balance in the MJO-diurnal cycle interaction is complicated by processes that should not be overlooked for a complete understanding and correctly numerical simulations of the intraseasonal SST variability. One such process is surface buoyancy flux. The net surface buoyancy flux (B_{net}) consists of three components, B_{heat} , B_{precip} , and B_{evap} , which are related to the surface heat flux (Q), precipitation (P) and evaporation (E), respectively:

$$B_{net} = B_{heat} + B_{precip} - B_{evap} = \alpha(C_p \rho)^{-1} Q + \beta S_0 (P - E). \quad (2)$$

In (2), α and β are the thermal and haline coefficients of expansion, and S_0 is the reference surface salinity. The surface buoyancy flux modulates the stability of the upper ocean and thereby the effectiveness of wind-driven mixing. Among the three component, B_{heat} is often the largest, B_{evap} the smallest and usually negligible. Positive buoyancy input into the upper ocean through precipitation is sometimes offset by negative buoyancy input due to surface cooling. If not, two significant consequences would emerge when the upper ocean receives net positive buoyancy input. One is the formation of a barrier layer (Lukas and Lindstrom 1991). Its existence tends to prevent nighttime entrainment cooling and make daytime solar absorption more efficient to increase the upper ocean and surface temperatures on longer (e.g., intraseasonal) timescales. An extreme case of the barrier layer is the thin surface ‘‘fresh lens’’ (Wijesekera et al. 1999). This very thin and stable isohaline layer (~ 2 m) allows a portion of solar radiation to penetrate through but meanwhile experiences directly surface cooling by evaporation and longwave radiation. The consequences are a weaker diurnal cycle in surface temperature and a temperature inversion with a cool surface layer above a warm layer. An episode of strong surface wind would erode this fresh lens and lead to a warming of the surface, the opposite to what expected without the fresh lens. Surface cooling on longer (e.g., intraseasonal) timescales would thus be reduced.

This complication in the MJO-diurnal cycle interaction due to surface buoyancy flux sensitively depends on the surface structure of the MJO (e.g., relative phases of its wind and rainfall anomalies). Because convective centers of the MJO usually coincide with its strong surface westerly wind, positive buoyancy input into the upper ocean due to precipitation indeed mostly offsets negative buoyancy input due to surface cooling during convective active phases of the MJO. But this offset would be misrepresented if heavy rain occurs outside the period of strong surface wind as in many numerical models. Biases in intraseasonal perturbations of SST due to errors in simulated MJO structures can be as large as those due to errors in other simulated MJO characteristics, such as its period, zonal wavelength, phase speed, and total surface heat flux (Fig. 2).

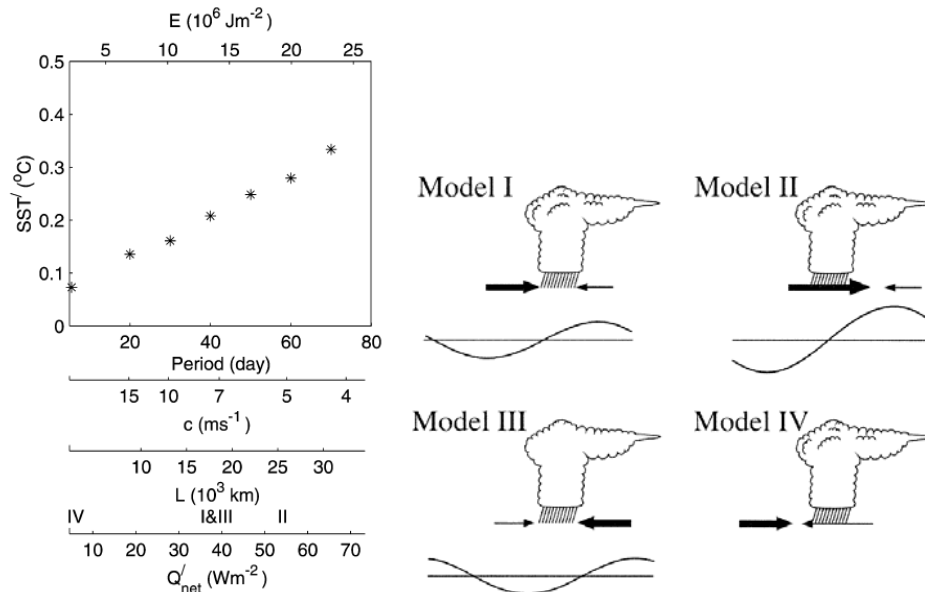


Figure 2 Left panel: Amplitudes of intraseasonal perturbations in SST as a function of the period, phase speed (c), zonal wavelength (L), and net surface heat flux (Q'_{net}) of the MJO and the total air-sea energy exchange throughout its life cycle (E). The standard conditions are $c = 5 \text{ m s}^{-1}$, $L = 26\,000 \text{ km}$, $Q'_{net} = 55 \text{ W m}^{-2}$, and period = 60 days. Marked at the abscissa of Q'_{net} are its values corresponding four idealized MJO structures sketched in the right panel. Right panel: Four idealized MJO structures. 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, which is also the location of maximum surface fluxes of latent and sensible heat from the ocean. SST perturbations are illustrated at the bottom of each structure except for model IV. From Zhang and Anderson (2003).

It has been shown that without explicitly resolving the diurnal cycle, numerical simulations may underestimate the amplitude of the intraseasonal perturbation in SST. Because of the fine structure of the upper ocean related to the barrier layer and fresh lens, a numerical ocean model would lose more than 50% of the diurnal SST variability if its vertical resolution in the upper ocean is less than 4 m. Most global coupled models do not resolve the diurnal cycle and their vertical resolutions are usually coarser than 4 m. If they reproduce the bulk properties of the MJO and simulate the right intraseasonal SST variation, it only means that they have more than one sources of errors that cancel each other.

Through its rectification onto intraseasonal variability in SST, the diurnal cycle may feed back to the MJO by helping organizing atmospheric deep convection over enhanced positive SST anomalies. Numerical

simulations have produced the MJO whose precipitation signals are stronger and more coherent in coupled models with the diurnal cycle explicitly resolved than without. In addition, strong diurnal fluctuations in SST during a convectively inactive phase of the MJO might be instrumental to shallow convection and precipitating cumulus congestus (Johnson et al. 1999) that may help moistening the lower troposphere to prepare the next convectively active phase. Such shallow precipitating convection can be essential to low-level large-scale moisture convergence that also facilitates the evolution of the MJO (Zhang and Mu 2005; Li et al 2008).

3. MJO and ENSO

Possible feedback between the MJO and ENSO has been a research subject since first suggested by Lau and Chan (1985). Modulations of the MJO by ENSO have been clearly observed. During ENSO warm events, the MJO propagates further eastward than it normally does into the central Pacific because of the anomalous warm sea surface there; the opposite occurs during ENSO cold events (e.g., Kindle and Phoebus 1995; Bergman et al 2001). MJO feedback to ENSO is, however, a controversial issue. This subsection focuses on two aspects of the cross-scale air-sea interaction associated with the MJO and ENSO: the role of the MJO as a source of stochastic forcing, and its excitation of intraseasonal oceanic Kelvin waves. Before these two topics are discussed, it may help to first brief the general issue of stochastic influences on ENSO.

Possible stochastic effects on ENSO was first suggested by Lau (1985). According to Hasselmann (1976), when a field z can be separated into its low-frequency component s and high-frequency component f , i.e., $z = s + f$, the variability of the low-frequency component can be driven by two types of dynamics,

$$ds/dt = W(s) + G(f), \quad (3a)$$

where W represents low-frequency dynamics, and G stochastic dynamics due to the high-frequency component, f . G can be written in a form of

$$G(f) = [M(s) + A]f, \quad (3b)$$

where M represents the feedback from s to f and A is independent of s . When $M = 0$, there is no feedback and G represents high-frequency dynamics alone, whose effect is commonly referred to as *additive*. When M is not zero, the inclusion of the feedback from s makes the stochastic process *multiplicative*. In the context of MJO-ENSO interaction, ENSO variability is represented by s , and MJO by f .

When the atmosphere-ocean coupled system is in an unstable dynamical regime, ENSO variability is mainly driven by the coupled dynamics W . Atmospheric stochastic component plays a secondary role by inducing additional irregularity in the ENSO variability. When the coupled system is in a neutral or stable dynamical regime, the coupled dynamics W alone is insufficient to sustain an ENSO variability and stochastic influences are essential to ENSO. The most fundamental controversy regarding the role of atmospheric stochastic perturbations is what dynamical regime the coupled systems is in (in other words, whether ENSO is primarily driven by stochastic processes) and how the dynamical regime may shift from one to the other on longer (decadal and inter-decadal) timescales. The following discussion is made under an assumption that the coupled system is slightly stable and atmospheric stochastic perturbations are essential to sustain the ENSO variability.

Coupled models of intermediate complexities, when tuned to be slightly stable and forced by additive stochastic surface wind derived from global reanalyses, can reproduce ENSO variability whose statistics and even timing resemble those observed (Zavala-Garay et al. 2005). In such a model, the MJO component of the stochastic wind is responsible for about 70% of the simulated ENSO variance. The large-scale coupled

dynamics (W in 3a) is also essential. Without the coupling, the MJO alone cannot induce any significant variability on interannual timescales. In other words, energy provided by the MJO, however weak, is amplified by the coupled dynamics and grow into observable ENSO variability. The role of the MJO in ENSO appears to be linear in that the interannual power associated with the MJO constitutes the essential part of the stochastic forcing, consistent to the original stochastic climate model of Hasselmann (1976). The key notion here is that the critical interannual stochastic power exists solely because of the MJO. In the absence of the low-frequency power, intraseasonal effects of the MJO on the ocean can be rectified on to interannual timescales through nonlinear processes, including the dependence of surface evaporative cooling on wind speed and Ekman convergence of westerly momentum (Kessler and Kleeman 2000; Shinoda and Hendon 2002).

The role of the MJO in ENSO simulated by a coupled global climate model (CGCM) can be diagnosed with a help of a coupled model of intermediate complexity and a global reanalysis (Kapur 2008). The diagnostic procedure is sketched in Fig. 3. A Thomas and Battisti (2000) version of the Cane-Zebiak model is used as the center tool, which is further modified by Zavala-Garay to allow daily stochastic forcing to be included. Atmospheric stochastic perturbations in surface wind are derived from both the reanalysis and CGCM simulation by removing wind component linearly coherent with ENSO SST. The MJO and non-MJO parts of the stochastic wind are then separated. The realism of the stochastic winds from the CGCM is assessed by comparing them to those from the reanalysis. ENSO statistics produced by the Cane-Zebiak-Zavala (CZZ) model with added stochastic wind from the reanalysis is closest to the observed when the model is tuned to be slightly stable. In this dynamical regime, ENSO statistics produced by the CZZ model forced by added stochastic wind from the CGCM are compared to those from observations, CGCM simulations, and CZZ simulations forced by added reanalysis stochastic wind to assess the role of stochastic variability in ENSO simulated by the CGCM.

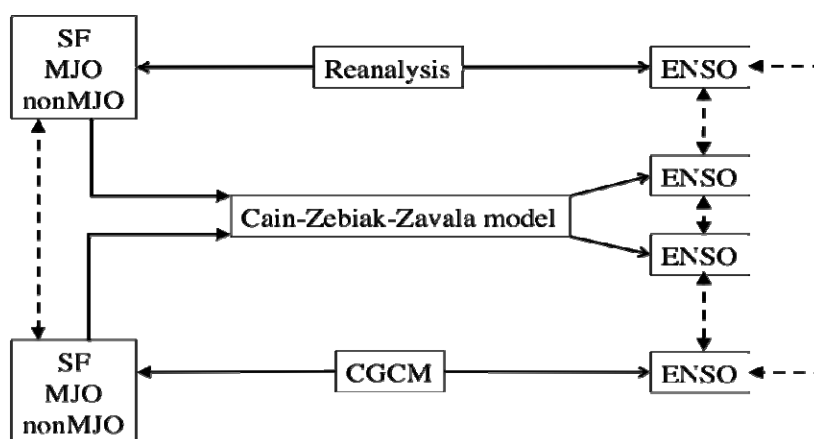


Figure 3 Sketch of a procedure for diagnosing the role of the MJO in ENSO simulated by a CGCM. Solid arrows indicate data flows, while dashed arrows mark comparisons and validations.

The above procedure demonstrates that the ENSO variability in the CGCM could be driven by stochastic forcing. Many ENSO statistics (e.g., PDF, spectrum, and seasonally dependent decorrelation of Niño 3 SST) seen in observations and CGCM simulations are reproduced by the CZZ model forced by stochastic wind from the CGCM. The major exceptions are the seasonal phase lock of ENSO warm events and the strength of cold events, which are attributed to the coupled dynamics. In the CZZ simulations with stochastic winds from both reanalysis and CGCM, the MJO components are dominant. Without the MJO component, the CZZ model reproduces the ENSO statistics poorly.

Observational evidence of the role played by the MJO in ENSO comes from two sources. One is the Pacific moored buoy network, which reveals energetic MJO events prior to almost all modern ENSO warm events (e.g., McPhaden 1999). The other is statistics based on global reanalysis products, which show significant lag correlation with ENSO SST lagging MJO activities by 6 – 12 month (Zhang and Gottschalck 2002). The phase lock of ENSO with the seasonal cycle and this lag correlation point to the importance of the MJO in boreal spring (Hendon et al. 2007).

The MJO plays a dominant role in stochastic forcing of ENSO for several reasons. Its zonal structure matches well the stochastic optimal of ENSO (Moore and Kleeman 1999). It is known to have strong interannual variability almost independent of ENSO (Slingo et al. 1999; Hendon et al. 1999). Its eastward slow propagation makes it much more effective in generating oceanic Kelvin waves than other atmospheric stochastic perturbations (Hendon et al. 1998). This leads to another topic: the role of intraseasonal oceanic Kelvin waves in connecting the MJO and ENSO.

The MJO may facilitate the growth of an ENSO warm event through several processes, all tending to reduce the zonal gradient of equatorial SST. In addition to cooling the western Pacific warm pool by its overlapping enhanced surface latent heat flux and reduction in insolation (Zhang and McPhaden 2000), the MJO excites oceanic downwelling Kelvin waves that advect the warm pool eastward (Kessler et al. 1995) and, when propagating into the eastern Pacific, deepen the thermocline, suppress entrainment cooling, and thus induce positive equatorial SST anomalies (Zhang 2001). These processes related to downwelling Kelvin waves do not necessarily cancel out by those related to upwelling Kelvin waves for two reasons. First, the surface zonal wind associated with the MJO is not symmetric (or sinusoidal). MJO westerly winds are far stronger than its easterly winds. A sequence of MJO events would result in a net cooling and eastward advection of the western Pacific warm pool (Shinoda and Hendon 2002; Kessler et al. 1995), and a net positive SST anomaly over the equatorial eastern Pacific (Zavala-Garay et al. 2008). Second, oceanic feedback to the MJO tends to make the latter stronger, propagate further eastward, and thus excite stronger oceanic Kelvin waves (Kessler et al. 1995). Such a feedback raises an important issue which is the last topic of this article: the multiplicative nature of ENSO stochastic forcing.

In the Hasselmann (1976) stochastic climate model, feedback from low- to high-frequency components, i.e., $M(s)$ in (3b), plays a distinct role to stabilize the amplitude of the low-frequency component. Multiplicativity in ENSO stochastic forcing may modify ENSO statistics only slightly (Perez et al. 2005), enhance ENSO amplitude significantly (Eisenman et al. 2005), or be an essential ingredient for ENSO instability (Jin et al. 2007), all depending on the model used. Our recent study on multiplicative stochastic MJO forcing to ENSO in the CZZ model suggests that almost all possible results can be obtained within a broad range of model parameter space. It is fair to say that the relative importance of additive vs. multiplicative stochastic forcing to ENSO remain a subject of research.

4. Concluding remarks

This article briefly reviews two examples of cross-scale air-sea interaction in the tropics involving the MJO. The takehome message is that the properties of the MJO is central to a complete understanding of these cross-scale air-sea interaction phenomena and to numerical simulating the correct physical processes involved. Recent studies have demonstrated that ENSO simulations are improved when the MJO is better reproduced by the atmospheric component of CGCMs (Wu et al. 2007; Neale et al. 2008). In general, improved understanding and simulation of the MJO will benefit our ability of understanding, simulating and predicting a wide range of tropical weather and climate phenomena.

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