A conceptual model for self-sustained active-break Indian summer monsoon

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[1] The Indian summer monsoon (ISM) exhibits a salient intraseasonal oscillation (ISO), and there have two ISO variability centers that are located at the equatorial Indian Ocean (EIO) and ISM regions, respectively. The prevalence of these two centers indicates that this active-break monsoon cycle has a prominent stationary oscillation. To understand this aspect of the ISM ISO, a two-box model is advanced to represent the ISO in these two variability centers. Linking by the local Hadley circulation, these two boxes can enhance each other’s variability, but air-sea interaction is shown to be necessary for producing a sustainable model ISO. Associated with a positive (negative) precipitation dipole anomaly, the reduced (enhanced) short wave radiation cools (warms) the sea surface and leads to an anomalous sea surface temperature dipole, which in turn stimulates development of a reversed precipitation dipole. This leads to a transition from one to the other phase of the monsoon cycle. This model demonstrates that even without the northward propagation, the air-sea interaction, in cooperation with the instability caused by the local Hadley circulation and frictional boundary layer moisture convergence, can produce a self-sustained ISM ISO. Citation: Liu, F., and B. Wang (2012), A conceptual model for self-sustained active-break Indian summer monsoon, Geophys. Res. Lett., 39, L20814, doi:10.1029/2012GL053663.

1. Introduction

[2] The Indian summer monsoon (ISM) experiences a prominent intraseasonal oscillation (ISO), characterized by 20–50 day fluctuations with “active” period of heavy rain interrupted by dry “breaks” [e.g., Krishnamurti and Bhalme, 1976; Gadgil, 2003]. The active-break monsoon not only affects the whole Indian Ocean and adjacent continental regions, but also influences remote tropics and North Pacific Ocean [Webster et al., 1998]. In contrast to the boreal winter Madden and Julian [1971] Oscillation (MJO) which, to some extent, is re-initiated by the globally circumnavigating upper-level divergent waves [Lorenc, 1984], the active-break ISM has been found to be a self-induced oscillatory system in recent observations [Wang et al., 2005], where the initiation of the ISO convection in the western equatorial Indian Ocean (EIO) is primarily a local process within the ISM system [Jiang and Li, 2005].

[3] During boreal summer the ISM trough and eastern EIO are two salient ISO variability centers (Figure 1a), in between which exists northward propagation. The oscillations between these two centers are basically 180 degrees out of phase (Figure 1b). Associated with the active-break precipitation cycle, the sea surface temperature (SST) also experiences prominent ISO over the Bay of Bengal (BoB) and EIO (Figure 1b). Meanwhile, positive precipitation anomaly leads to ocean cooling, and warm SST anomaly leads positive precipitation anomaly (Figure 1c), which suggests that the air-sea interaction may be an important process in the self-induction of the active-break ISM.

[4] Over the Indian Ocean and Western Pacific Ocean, the air-sea interaction favors northward propagation of the boreal summer ISO [Kemball-Cook and Wang, 2001; Fu et al., 2003; Rajendran and Kitoh, 2006], which tends to suppress the unstable growth and reduce the oscillation’s period [Bellon et al., 2008]. However, these works didn’t explain why these two strong ISO variability centers located at the eastern EIO and BoB can sustain a stationary seesaw oscillation on the intraseasonal time scale.

[5] Despite the primary importance of the active-break ISM and years of research, a generally accepted explanation for its essential mechanism remains elusive. We present a conceptual model to demonstrate one of the basic mechanisms behind this self-induction of the active-break ISM.

2. Conceptual Model for the Self-Sustained Active-Break ISM

2.1. Air-Sea Interaction Model

[6] The framework, based on a coupled atmosphere-ocean model derived for the warm pool climate system [Wang and Xie, 1998], mainly captures the strong air-sea interaction in the ISM ISO (Figure 1c): 1) For perturbations, positive SST systematically induces positive precipitation, i.e., the SST can affect the pressure through surface evaporation [Zebiak, 1986] and longwave radiation [Davey and Gill, 1987; Lindzen and Nigam, 1987]. 2) The SST is also affected by the precipitation (via cloudiness) through changing downward short wave radiation (SWR) [Wang and Xie, 1998; Fu et al., 2003]. We assumed that in the warm pool region, the oceanic dynamical effect on the SST is negligibly small compared to the effects of surface heat flux exchange following the formulation of Wang and Xie [1998]. 3) To represent the role of the planetary boundary layer, we use the Lindzen-Nigam model [Lindzen and Nigam, 1987; Neelin, 1989; Wang and Li, 1993]

\[ E\vec{V}_a + y\beta \hat{k} \times \vec{V}_a = -\nabla \phi + G\nabla T. \]
\[ \varepsilon u - \beta y v = -\phi_x, \]
\[ \beta y u = -\phi_y, \]
\[ \phi_t + \mu \phi = -C_0 D + C_0^2 (D - rv) - I_s (\alpha_E + \alpha_L) T, \]
\[ T_t + \mu \phi T = D_{rad} (D - rv). \]

\[ u \text{ and } v \text{ are lower-tropospheric velocities, and } D = u_x + v_y \text{ the divergence.} \]
\[ J = RL_C \Delta p q_L / (2p_2 C_2 C_0), \]
\[ r = b \Delta p q (q_B - q_L) / (\Delta p q_L), \]
\[ I_s = R g (C_2 / p_2), \]
\[ D_{rad} = 0.622 L_c q_L (1 - A)^{2} / (p_0 C_0 h). \]

\[ \text{is the oceanic mixed-layer depth, and } b = 1.5 \text{ plane}\]
\[ \text{is the surface albedo, } \]
\[ \text{the middle-tropospheric pressure, lower-tropospheric and boundary layer pressure depths, respectively, } \]
\[ b \text{ the available moisture ratio determining how much the boundary layer moisture can be used for the diabatic heating. } \]
\[ q_B \text{ and } q_L \text{ stand for the mean specific humidity at the boundary layer (1000 hPa through 900 hPa) and the lower troposphere (900 hPa through 500 hPa), respectively [Wang and Xie, 1988].} \]
\[ A \text{ is the surface albedo, } \]
\[ S_0 \text{ the sea surface solar radiative flux under clear sky, } \]
\[ C \text{ the perturbation cloud cover, which is assumed to be proportional to the perturbation precipitation with a coefficient } \gamma. \]

\[ 8 \text{ The constants include the specific gas constant } R = 287 \text{ J} \cdot \text{kg}^{-1} \cdot \text{K}^{-1}, \text{ the specific heat at constant pressure } C_p = 1004 \text{ J} \cdot \text{kg}^{-1} \cdot \text{K}^{-1}, \text{ the water heat capacity } C_w = 4186 \text{ J} \cdot \text{kg}^{-1} \cdot \text{K}^{-1}, \text{ the latent heat of condensation } L_C = 2.5 \times 10^6 \text{ J} \cdot \text{kg}^{-1}, \text{ the water density } \rho_0 = 1.0 \times 10^3 \text{ kg} \cdot \text{m}^{-3}, \text{ the gravity acceleration } g = 9.8 \text{ m} \cdot \text{s}^{-2}, \text{ and the meridional Coriolis parameter variation } \beta = 23 \times 10^{-11} \text{ m}^{-1} \cdot \text{K}^{-1}. \]

\[ 9 \text{ The diabatic heating } Q \text{ comes from the moisture convergence of the lower troposphere and frictional boundary layer, respectively, i.e., } \]
\[ Q = -L_C q L / \Delta p q_L (q_B - q_L) \text{ with } \Delta p. \]

\[ \text{The evaporation coefficient } \alpha_E = p_0 C_2 L K_g U, \text{ where } \]
\[ \alpha_L = 1.2 \text{ kg} \cdot \text{m}^{-3}, \text{ the surface air density, } C_E = 1.5 \times 10^{-3} \text{ the moisture transfer coefficient, and } K_g = 8.9 \times 10^{-4} \text{ K}^{-1}. \]

\[ 10 \text{ The longwave radiation coefficient } \alpha_L = \mu C_p \Delta p / g [Wang and Li, 1993]. \]
\[ h \text{ is the oceanic mixed-layer depth, and both boxes are first assumed to have the same depth of } 20 \text{ m.} \]
\[ \text{Since the precipitation is simply } Q \Delta p / (L_g h) [Wang, 1988], \]
\[ \text{the typical value } \gamma = 45.4 \text{ kg} \cdot \text{s}^{-1} \text{ means that an anomalous} \]

\[ y \text{ is the meridional displacement, } \vec{V}_B \text{ the boundary-layer horizontal wind, } \phi \text{ the geopotential anomaly, } T \text{ the SST, } E \text{ the boundary layer friction, } \beta \text{ the meridional variation of Coriolis parameter, and } G \text{ the forcing parameter associated with the SST gradient. Then the Ekman pumping at the top of the boundary layer is [Xie and Wang, 1996]} \]
\[ w = (d_1 \partial_{xx} + d_1 \partial_{yy} + d_2 \partial_z + d_3 \partial_y) \phi, \]
\[ \{d_1, d_2, d_3\} = \{E(E^2 + \beta^2 y^2) - \beta(E^2 - \beta^2 y^2) / (E^2 + \beta^2 y^2)^2, -2E \beta^2 y / (E^2 + \beta^2 y^2)^2\}. \]
precipitation of 1 mm day\(^{-1}\) may result in an increase in total cloudiness by one-fifth.

2.2. Two-Box Model

[11] Two boxes are located on the convection centers at the monsoon region (15°N, suffix “m”) and the equatorial region (Equator, suffix “e”). These two boxes are characterized by different atmospheric processes, i.e., Kelvin wave in the EIO and Rossby wave in the BoB, and by different oceanic features, i.e., a shallower mixed layer in the BoB than that in the EIO [Bellon et al., 2008].

[12] As a planetary-scale system, we neglect the zonal variation near the convection center and the dynamic instability mainly comes from the frictional boundary layer convergence [Wang, 1988], then (3) becomes independent of \(x\). Assuming that perturbations vanish at the lateral boundary, i.e., \(\Phi_{15S} = \Phi_{50N} = 0\), the linear interpolation yields \(\Phi_{y=7.5S} = \Phi_{e/2}\), \(\Phi_{y=7.5N} = (\Phi_m + \Phi_e)/2\), and \(\Phi_{y=22.5N} = \Phi_{m/2}\). Then the central difference method gives \(\Phi_{my} = -\Phi_e/(4\Delta y)\), \(\Phi_{myy} = (\Phi_e - 2\Phi_m)/(2\Delta y)\), \(\Phi_{ey} = \Phi_{myy}/(4\Delta y)\), and \(\Phi_{eyy} = (\Phi_m - 2\Phi_e)/(2\Delta y)^2\), where the meridional step \(\Delta y = 7.5°\). With the momentum equations in (3) we can obtain the divergence \(D\) in two different boxes, thus the divergence and Ekman pumping in these two boxes are

\[
\{D_m, D_e, w_m, w_e\} = \left\{ \begin{array}{c}
\frac{-e}{2\Delta y^2} \Phi_y - \frac{e}{2\Delta x^2} \Phi_x - \frac{e}{2\Delta y^2} \Phi_{eyy} (\Phi_e - 2\Phi_m), 0, \\
-\frac{d_1}{\Delta x^2} \Phi_m + \left( \frac{d_1}{2\Delta x^2} - \frac{d_1}{4\Delta y} \right) \Phi_e - \frac{d_1}{\Delta x^2} \Phi_x \\
+ \left( \frac{d_1}{2\Delta x^2} + \frac{d_3}{4\Delta y} \right) \Phi_e \end{array} \right. \}
\]

which can be calculated analytically as a linear system.

3. Simulated Self-Induction of the Active-Break ISM

[13] Let’s begin with a single box model. This can be done by removing one of the two boxes. The atmospheric only model can be further obtained by setting \(D_{rad} = 0\). Under current parameters, the single equatorial or monsoon box model alone produces a damping mode for the precipitation (Figure 2a). Figure 2b shows that the coupling of these two boxes produces a growing (with a growth rate of 0.06 day\(^{-1}\)) precipitation dipole. The reason is that the equatorial and monsoon precipitation can enhance each other through the local Hadley circulation, i.e., the positive equatorial precipitation produces downward motion and negative precipitation anomaly in the monsoon region and vice versa. As a growing system, Figure 2b shows a monotonic evolution, and there is no oscillation. Without the air-sea interaction, the atmospheric process can hardly explain the ISO over the ISM.
When the air-sea interaction is included, the isolated equatorial or monsoon box still produces a damping mode, but a damped oscillation emerges (Figure 2c), for example, the EIO precipitation begins to oscillate with a period of 45 days. It is interesting that the two-box air-sea interaction model can successfully simulate the ISO-like oscillation of the ISM (Figure 2d). Like in observation (Figure 1b), both the monsoon and equatorial precipitations experience active/break phases. Meanwhile, the interaction between the equatorial and monsoon precipitations not only enhances each other, but also reduces the oscillation’s period to 35 days. The air-sea interaction plays an essential role in sustaining this oscillation. The leading positive anomalous SST warms the lower troposphere and prepares moisture for the deep convection [Johnson et al., 1999]. The decrease of the SWR associated with the strong precipitation tends to cool the sea surface [Fu et al., 2003], and produces a negative anomalous SST that lags the precipitation.

The solutions obtained above are parameter-dependent. For example, a warm SST or efficient boundary layer effect, i.e., a large parameter $b$, will enhance the instability (Figure 3a), because the atmosphere in these boxes can become unstable owing to the frictional boundary layer convergence feedback [Wang, 1988; Xie and Wang, 1996]. As a linear model, the growth rate increases linearly with increasing $b$; and the period of the unstable mode shortens when $b$ increases (Figure 3a). Sensitivity experiments also show that the strong mean surface wind will enhance the evaporation as well as the air-sea interaction, which tends to reduce the oscillation’s period.

Most importantly, the modeled results strongly depend on the mixed-layer depth, i.e., the efficiency of the SWR effect, $D_{rad}$. In Figure 3b, a reduced mixed-layer depth would amplify the SWR effects, thus reduce the modeled oscillation period; in this case, air-sea interaction does provide a negative feedback to maintain the ISO [Bellon et al., 2008]. In the coupled general circulation model experiment by Fu et al. [2003], the oceanic mixed-layer depth is found shallower (15 m) in the monsoon region and deeper (35 m) in the equatorial region. With these parameter values, our model yields an oscillation with a period of 36 days (Figure 3b).

4. Concluding Remarks

This box model describes the physical processes behind the self-sustained ISM ISO. The transition from a break to active phase consists of two phases as illustrated by Figure 4 and described as follows. First, in the break phase of ISM, a positive (negative) SST anomaly exists in the eastern EIO (BoB) due to delayed oceanic response to the atmospheric forcing. This dipole SST anomaly will warm (cool) the lower troposphere and result in low (high) pressure in the eastern EIO (BoB), thus developing a precipitation dipole (enhanced over the EIO and suppressed over the BoB) due to the instability caused by the frictional boundary layer convergence and the connection by local Hadley circulation between the equatorial and monsoon regions. Secondly, the positive (negative) cloud anomaly associated with the positive (negative) precipitation perturbation produces negative (positive) SWR anomaly, which will cool (warm) the sea surface in the EIO (BoB), thus a reversed SST dipole anomaly appears in the following active phase of ISM. This model reveals that even without the northward propagation, the air-sea interaction, in cooperation with the instability caused by the local Hadley circulation and
frictional boundary layer moisture convergence, can produce a self-sustained ISO over the EIO and BoB.

[15] In this work we neglected some nonlinear processes. For example, the mixed-layer depth can be affected by the changes of SWR or SST [Bellon et al., 2008], and it in turn affects the SST by changing $D_{rad}$. Further, the anomalous wind also affects the mean flow in the western EIO, which might be important for the reinitiation of the ISO [Jiang and Li, 2005]. Over the BoB and eastern EIO, the relationship between the anomalous and mean winds is complicated, and we hope to discuss the consequence of these nonlinear processes in future. Although the stationary oscillation exists in the ISM, the northward propagation is also important [Sikka and Gadgil, 1980; Lawrence and Webster, 2002; Rajendran and Kitoh, 2006; Bellon et al., 2008] because it would more accurately predict the active-break ISM cycle. These deficiencies call for further improvement of this simple theory.

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