

## Comments on "An Air-Sea Interaction Model of Intraseasonal Oscillation in the Tropics"\*

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### 1. Introduction

Emanuel's (1987) model of intraseasonal oscillation posited that the waves producing 30–60 day oscillation arise from an air–sea interaction instability driven by wind-dependent surface fluxes of moist entropy. The conceptual notion is illustrated in his Fig. 1 and may be summarized as follows.

The mean surface winds in the model are assumed to be easterlies; thus, the perturbation easterlies will result in an anomalously large flux of latent heat from the sea surface, while perturbation westerlies will be associated with negative anomalies of surface heat flux (for convenience this will be referred to as Assumption A). The moist convection in the model is assumed to immediately redistribute the heat obtained from the sea surface but does not act as a heat source in or of itself (henceforth referred to as Assumption B). With these two assumptions, it follows that the resulting tropospheric warming will lead the vertical motion (or surface convergence) of the wave by a quarter of wavelength, leading to wave growth and eastward propagation.

In this note I wish to make some comments on the foregoing conceptual notion in an attempt to stimulate further research.

### 2. The representation of evaporation heating

In Emanuel's model, the east–west asymmetries in surface latent heat fluxes play a key role in determining the phase propagation and growth of the 30–60 day waves. In a linear dynamic framework, easterly (westerly) anomaly may induce positive (negative) anomaly of surface heat fluxes only when the mean surface flow is easterly with a *significant* strength (say,  $2 \text{ m s}^{-1}$ ). If the mean surface wind speed is near zero, there will be

negligible difference in the surface heat fluxes between easterly and westerly anomalies; accordingly, the heating representation of the model will not apply.

It is true that the global mean surface winds in equatorial region are easterlies. However, it is important to note the remarkable longitudinal variations in the mean surface winds in the equatorial zone. We have computed 80-year (1900–79) monthly mean surface wind components on a  $2^\circ$  longitude by  $20^\circ$  latitude grid running from west to east centered on the equator, using data from comprehensive ocean–atmosphere dataset (COADS) (Woodruff et al. 1987). The continental regions of equatorial Africa (between  $10^\circ$  and  $40^\circ\text{E}$ ) and South America (between  $80^\circ$  and  $50^\circ\text{W}$ ) were excluded. The reason we have concentrated our attention to the equatorial region between  $10^\circ\text{N}$  and  $10^\circ\text{S}$  is that the tropical 30–60 day oscillations are confined meridionally to low latitudes with maximum amplitude located along the equator (e.g., Madden and Julian 1972; Lorenc 1984; Weickmann et al. 1985). The monthly mean surface westerlies are contoured in Fig. 1 as functions of time and longitude. Strong easterlies prevail over the central and eastern Pacific and Atlantic Oceans with a maximum speed close to  $6 \text{ m s}^{-1}$  occurring near  $145^\circ\text{W}$  during northern winter. On the other hand, weak westerlies are found over the Indian Ocean with a maximum speed in excess of  $2 \text{ m s}^{-1}$  occurring in the central Indian Ocean during northern summer. In the equatorial regions of the western Pacific and maritime continent between  $100^\circ$  and  $160^\circ\text{E}$ , the monthly mean surface winds tend to be very weak and to reverse their directions in the region between  $110^\circ$  and  $140^\circ\text{E}$  from westerly in northern winter to easterly in northern summer. In his study of pressure–wind relationship, Lettau (1974) previously noticed the existence of seasonally persistent westerly winds at 900 mb in the vicinity of the equator from  $40^\circ$  to  $170^\circ\text{E}$ .

Figure 2 shows the longitudinal variations of annual and seasonal means of surface westerlies averaged over the equatorial zone between  $10^\circ\text{N}$  and  $10^\circ\text{S}$ . Overall, the mean surface winds over the equatorial ocean of the eastern hemisphere are very weak. The annual mean surface westerlies averaged between  $0^\circ$  and

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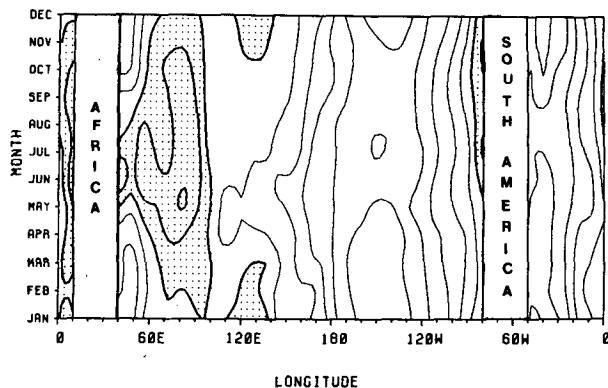


FIG. 1. Longitude-time section of monthly mean surface westerly component (contour interval:  $1 \text{ m s}^{-1}$ ) along the equator ( $10^\circ\text{N}$ – $10^\circ\text{S}$ ). Regions of westerlies are denoted by meshed areas. The monthly means are computed from 80-year (1900–79) COADS data using  $8^\circ$ -longitude running mean. The blank areas between  $10^\circ$  and  $40^\circ\text{E}$  and between  $80^\circ$  and  $50^\circ\text{W}$  represent African and South American continents, respectively.

$160^\circ\text{E}$  is merely  $-0.25 \text{ m s}^{-1}$ . I wish to point out that the longitudinal domain in which significant equatorial easterlies are nonexistent has a near-hemispheric scale. This poses a severe limitation on the applicability of the model's heating representation for 30–60 day waves which have typical zonal scales of wavenumber one or two.

The absence of significant easterlies over the Indian and western Pacific Oceans also implies that the geographic locations favorable for supporting the east–west asymmetries in evaporation anomalies and for wave amplification are the central-eastern Pacific and Atlantic oceans. Yet observations indicate that the anomalous circulation as well as the cloudiness associated with 30–60 day waves exhibit the largest amplitude over the Indian and western Pacific while decay as the waves transverse the eastern Pacific and Atlantic Oceans (e.g., Madden and Julian 1972; Krishnamurti et al. 1985; Weickmann et al. 1985; Murakami et al. 1986; Knutson et al. 1986; Lau and Chan 1988).

The longitudinal variability of observed intraseasonal oscillation may be suggestive of their relation to sea surface temperature (SST). Lau and Chan (1986) pointed out that the oscillation in OLR tends to be confined to the equatorial region of the Indian and western Pacific Oceans in non-El Niño year and to expand eastward into the central and eastern Pacific Ocean during El Niño–Southern Oscillation (ENSO) events, suggesting a correlation between high SST and enhancement of the oscillation. We should note that during the warm episodes in the central and eastern tropical Pacific the spread of positive SST anomaly coincides with the intensification of westerly anomaly (Rasmusson and Carpenter 1982). During 1982 to 1983 ENSO event the normal easterlies in the central Pacific were observed to become westerlies (Harrison

and Cane 1984). Therefore, the amplification of the 30–60 day waves tends to link to a warm ocean surface rather than to the zonal asymmetries in surface heat fluxes which depend on the existence of mean surface easterlies.

Besides the longitudinal variation of the mean surface winds, the SST also exhibits remarkable longitudinal variation which is not considered in the heating representation of the air–sea interaction model. In general, the regions of strong surface easterlies along the equator are accompanied by low SST due to wind-induced upwelling. Although the east–west asymmetries in surface heat fluxes are favored in these regions, one may seriously question the means by which the heat could be immediately redistributed aloft by convection as supposed in Assumption B. This is because convection tends to occur over warm ocean surface where SST exceeds about  $27.5^\circ\text{C}$  (Graham and Barnett 1987). In regions of strong surface easterlies the SST is low (e.g., annual mean SST over the equatorial eastern Pacific and Atlantic oceans is below this threshold value) and the troposphere is stable, so that convection is substantially suppressed. In view of the remarkable longitudinal variations of both mean surface winds and SST, the heating representation, in which the heating rate is solely proportional to perturbation easterlies, appears to be oversimplified.

In idealized zonally symmetric GCMs, equatorial mean surface winds are easterlies, and the Assumption A is obviously applicable to these simplified model atmospheres. However, the spectral peak corresponding to atmospheric intraseasonal oscillation in GCM simulations does not, in general, depend on the existence of mean surface easterlies and the asymmetries in surface heat fluxes. For instance, the climatological mean surface winds simulated by a more realistic GCM with land–sea contrast and topography resemble the observed tropical mean surface winds (Lau 1985). The

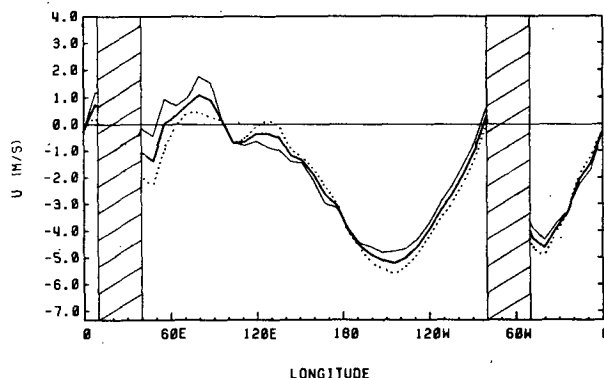


FIG. 2. Longitudinal variations of annual mean (thick solid), northern summer (May to October) mean (thin solid), and northern winter (November to April) mean (dotted) of the surface westerlies along the equator between  $10^\circ\text{N}$  and  $10^\circ\text{S}$ . Blank regions denote Africa and South American continents, as in Fig. 1.

long-term integration using this model clearly displays characteristics of intraseasonal oscillation (Lau and Lau 1986). Even a model with a zero-heating capacity ("swamp") lower boundary in which the wind-evaporation feedback is absent, is also able to simulate intraseasonal oscillation mode, which are essentially similar in their structure to those obtained from sophisticated GCM (Neelin et al. 1987). It seems that the way by which evaporation heating affects low frequency motion may differ from what is speculated by assumptions A and B. Further investigations are needed in this regard.

**3. The energetics**

Based on assumptions A and B, Emanuel inferred that the resulting temperature anomaly leads the surface convergence by one quarter cycle (his Fig. 1). He also stated that: "Since the temperature anomalies will have a component which is in phase with vertical velocity, potential energy is converted to mechanical energy, which can lead to the growth of the wave."

It should be first pointed out that if the temperature anomaly leads the surface convergence by one quarter cycle, there will be no kinetic energy generated (i.e., the covariances  $w'T'$  or  $u'\partial T'/\partial x$  averaged over a wavelength, will exactly vanish), thus no wave growth is possible.

For inviscid two-dimensional motion ( $v = 0$ ) in the equatorial zonal plane, the zonal momentum and thermodynamic equations of Emanuel's model [his equations (27) and (29)] may be reduced to, for easterly basic flow, the form:

$$\left(\frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial x}\right)u = \frac{\partial T}{\partial x}, \tag{1a}$$

$$\left(\frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial x}\right)T = -Au, \tag{1b}$$

where  $\bar{u}$  denotes the mean easterly wind,  $A > 0$  is a coefficient measuring the intensity of surface heat exchange. Assuming perturbation zonal wind and temperature of the form:

$$(u, T) = \text{Re}(U, \Theta)e^{ik[x - (\bar{u} + c)t]}, \tag{2}$$

where "Re" means taking real part of a complex. From (1a) and (1b), one obtains:

$$\Theta = -cU \tag{3}$$

and

$$c = \left(\frac{A}{k}\right)^{1/2} e^{i\pi/4}. \tag{4}$$

In (4) only equatorially trapped eastward traveling wave was considered. Equation (3) indicates that for a growing air-sea interaction mode, warming lags easterly anomaly or leads surface convergence by a phase angle  $\pi/4$ , rather than a quarter cycle,  $\pi/2$ .

Multiplying (1b) by  $T$  yields eddy available potential energy equation in which the rate of generation of available potential energy is given by

$$-AuT = \frac{\sqrt{2}}{2} |c| U^2 A e^{2kc t} \cos\gamma(\cos\gamma - \sin\gamma), \tag{5}$$

where  $\gamma = k[x - (\bar{u} + c_r)t]$  is the phase angle of zonal wind anomaly. Because the model assumed an exact compensation of moisture convergence-induced heating by adiabatic cooling (as implied in Assumption B), there is no conversion between the eddy available potential energy and eddy kinetic energy. Figure 3 illustrates the phase relationship between the wave energy generation and zonal wind anomalies. The regions of maximum generation is nearly symmetric about the surface convergence and nearly coincides with the maximum westerly and easterly anomalies (the phase difference is one-sixteenth of wavelength). Note that over the surface convergence phase there is no net generation of wave energy. Also note that the upward motion in this model lags maximum heating by a quarter of wavelength. This is uncommon in the tropics because the heating and vertical velocity are always strongly correlated for thermally direct circulation.

Let us now examine the energetics of the GCM counterpart of the atmospheric intraseasonal oscillation. The model to be analyzed is the same GCM used by Neelin et al. (1987) to show the importance of the evaporation-wind feedback to producing spectral peak in the intraseasonal range. By means of a specifically designed technique of composition, Lau et al. (1988) presented detailed vertical distributions of pressure velocity ( $w'$ ), diabatic heating rate ( $Q'$ ), and temperature anomaly ( $T'$ ) associated with intraseasonal oscillation mode. The composite scheme emphasizes these periods of strong velocity potential fluctuations with zonal scales of wavenumbers one to three and yields peak-to-peak amplitude of the fluctuations.

The covariance between perturbation temperature and diabatic heating  $\langle Q'T' \rangle$ , is computed for the com-

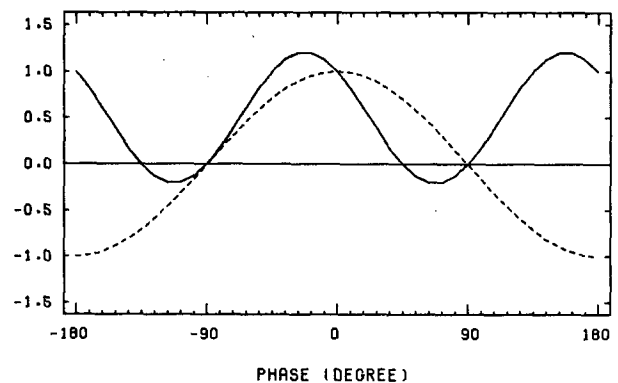


FIG. 3. Illustration of the phase relationship between zonal wind anomaly (dashed curve) and the generation of eddy available potential energy (solid curve) of the air-sea interaction model.

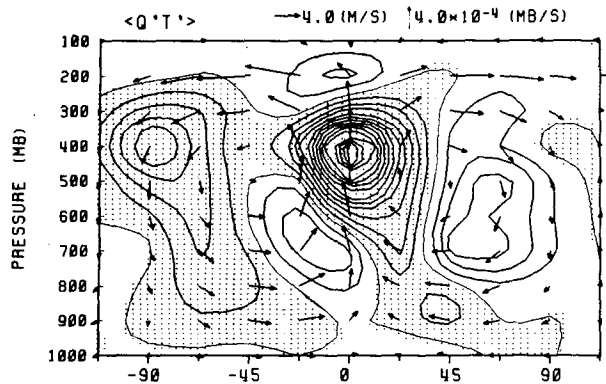


FIG. 4. Longitude-pressure distributions of covariance  $\langle Q'T' \rangle$  (contour interval:  $2.0 \times 10^{-6} \text{ deg}^2 \text{ C day}^{-1}$ ) along the equator. The data used for computation are subtracted from composites of intraseasonal oscillation modes in the tropics of an idealized GCM with zonally symmetric climate (Lau et al. 1988). The vectors indicate wind anomalies in the plane. Regions of positive values are shaded.

posite intraseasonal mode and contoured in Fig. 4 along with the circulation pattern in the equatorial zonal plane. The quantity  $\langle Q'T' \rangle$  is linearly proportional to the rate of generation of eddy available potential energy. Figure 4 shows that the majority of the available potential energy is generated in upper troposphere between 300 and 500 mb over the *surface convergence phase* of the wave. Another noticeable feature is the *asymmetry* in energy generation between westerly and easterly phases. Moreover, the eddy kinetic energy is also converted from eddy available potential energy in the surface convergence phase (figure is not shown). This is because the *diabatic heating is nearly in phase with the rising motion*. All these features in energetics are qualitatively different from the predictions of the air-sea interaction model. The GCM results suggest that the condensational heating associated with moisture convergence plays a more important part than that of evaporation anomalies. Some previous observational studies also pointed out that the 30–50 day wave motion is a thermally direct circulation (Murakami et al. 1984; Krishnamurti et al. 1985).

Vertical integration of moisture conservation equation for perturbation motion yields (Stevens and Lindzen 1978; Gill 1982):

$$P' = - \int_0^{p_0} \nabla_2 \cdot (\bar{q} \mathbf{V}'_2) \frac{dp}{g} + E' \quad (6)$$

where the temporal change of perturbation moisture content has been neglected. Equation (6) states that perturbation precipitation rate,  $P'$ , is approximately balanced by the sum of the perturbation moisture convergence from a unit column of atmosphere and local evaporation rate,  $E'$ . The anomalous condensational heating is thus related to circulation-dependent moisture convergence and local surface evaporation. There are two possible interactive processes involved in the

large scale moist wave dynamics, which may be described as precipitation-convergence feedback and evaporation-wind feedback. The scale analysis shows that a perturbation precipitation rate of about  $3 \text{ mm day}^{-1}$  is needed if the baroclinic structure of the observed 30–60 day wave is to be maintained (Wang 1988). The estimations based on observation and GCM simulation both indicate that the precipitation rate associated with intraseasonal mode is about that amount and is an order-of-magnitude greater than the perturbation evaporation rate, especially between  $10^\circ \text{S}$  and  $10^\circ \text{N}$ . The condensational heating appears to be dominantly supported by the moisture convergence rather than the local evaporation anomaly. Since the energy generation and conversion associated with these two mechanisms are different from each other, further observational diagnosis will help to determine the relative importance of the two feedback mechanisms for the development of the intraseasonal modes.

#### 4. Summary

In the air-sea interaction model for intraseasonal oscillation (Emanuel 1987), the amplification and phase propagation of unstable waves are controlled by the east-west asymmetries in circulation-dependent surface heat fluxes. The interpretation of the intraseasonal oscillation is based on the assumption that the mean surface winds are easterlies with significant strength. The observed mean surface winds, however, exhibit remarkable longitudinal variations. Significant surface easterlies prevail only over the central-eastern Pacific and Atlantic oceans. The annual mean of the surface westerlies averaged between  $0^\circ$  and  $160^\circ \text{E}$  is near zero ( $-0.25 \text{ m s}^{-1}$ ). The absence of meaningful surface easterlies in a near-hemispheric scale is expected to severely limit the model's applicability to observed tropical atmosphere. Furthermore, observed 30–60 day waves are found to amplify over the Indian and western Pacific oceans but the absence of mean surface easterlies implies that the evaporation-wind feedback is not favored in these regions. Over the western hemisphere equatorial oceans strong easterlies could support east-west asymmetries in surface heat fluxes; but the sea surface temperature is low due to wind-induced upwelling and the troposphere is stable, thus it might be inadequate to assume that convections immediately redistribute anomalous heat fluxes aloft and cause the 30–60 day wave to develop. In fact the wave circulation exhibits significant weakening as these waves travel across the eastern Pacific and Atlantic oceans, regardless of the evaporation-wind feedback being favored there.

Although Neelin et al. (1987) found that evaporation-wind feedback is important to intraseasonal oscillation in a GCM with zonally symmetric climate (the heating representation is applicable to this idealized model tropics), they also noticed that the existence of the spectral peak on intraseasonal time scale does

not depend on this mechanism because in a "swamp" model which rolls out evaporation-wind feedback, the intraseasonal oscillation still exists and the wave structure is essentially the same. Further analysis of the model used by Neelin et al. (1987) shows that the eddy available potential energy is mainly generated and then converted to mechanical energy in the surface convergence phase. The air-sea interaction model, however, predicts that no available potential energy is generated in the surface convergence phase and no kinematic energy is converted from eddy available potential energy. Future observational diagnosis is suggested to determine the relative importance of the precipitation-convergence feedback and the evaporation-wind feedback in supporting the tropical intraseasonal oscillation.

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