Some Dynamic Aspects of the Equatorial Intraseasonal Oscillations

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ABSTRACT

The slowly eastward-moving equatorial convection and circulation anomaly is a major mode of the tropical intraseasonal oscillations. The dynamics of this mode may be understood in terms of moist equatorial wave dynamics. A linear semi-geostrophic model on equatorial Beta-plane is used to study the behavior of equatorial low-frequency motions.

The unstable interaction of boundary layer frictional moisture convergence with condensational heating could generate efficiently eddy available potential energy for moist Kelvin mode but not for long Rossby modes. The growing mode is rooted in a moist Kelvin wave but modified through coupling with a long Rossby wave of the lowest meridional index.

The horizontal mode coupling makes the growing mode have a horizontal structure bearing similarity to both Kelvin and Rossby modes. It favors the amplification of long planetary scales and slows down the eastward movement. It also suppresses unrealistically fast growth of the uncoupled Kelvin waves by creating substantial meridional flows which induce kinetic energy destruction.

The model results also demonstrate that when maximum SST moves from the equator to 7.5^{o}N, the growth rate of the unstable wave is significantly reduced, suggesting that the annual march of the "thermal equator" and associated convective heating is likely responsible for annual variations of the equatorial intraseasonal wave activities.

1. INTRODUCTION

The slowly eastward moving equatorial circulation anomaly first described by Madden and Julian (1972) is believed to be a dominant mode of tropical intraseasonal variations. In the eastern hemisphere, the circulation anomalies are accompanied by convection anomalies, migrating due east at a speed of about 5 ms^{-1} (e.g., Lau and Chan, 1985; Weickmann et al., 1985; Murakami et al., 1986). After crossing the date line, the convection anomalies decay and emanate away from the equator towards North America and the southeastern Pacific (Rui and WAng, 1989), while the circulation anomalies can transverse the equatorial western hemisphere at a much faster speed of about 15-20 ms^{-1}(Knutson et al., 1986).

Numerical studies which rule out topographic and thermal land-sea contrast effects demonstrate that the precipitational heating interacting with equatorial wave motions may maintain long lasting transient planetary scale disturbances with eastward propagation speed of 10-20 ms^{-1} (Hayashi and Sumi, 1986; Lau and Peng, 1987; Swinbank et al., 1987; Lau et al., 1988).

Attempts have been made to understand the oscillations in terms of moist equatorial wave dynamics. A number of recent theoretical model studies have provided useful insights into moist Kelvin wave dynamics (e.g., Lau and Shen, 1988; Chang and Lim, 1988; Wang, 1988). Yet these analyses are confined to two-dimensional motion in the equatorial zonal plane. In this paper, we augment the moist Kelvin wave model by including meridional wind component, boundary layer frictional effect, and latitude-dependent SST for the basic state. It allows investigation of the impacts of horizontal-mode coupling and annual march of SST on low frequency equatorial wave dynamics.

2. SIMPLIFIED DYNAMIC FRAMEWORK

Consider perturbation motion on an equatorial Beta-plane. The primitive equation in p-coordinates are

$$\partial_t u-\beta yu=-\partial_x\phi+F_x \quad , \tag{1a}$$

$$\partial_t v+\beta yu=-\partial_y\phi+F_y \quad , \tag{1b}$$

$$\partial_x u+\partial_y v+\partial_p\omega=0 \quad , \tag{1c}$$

$$\partial_t(\partial_p\phi)+S(p)\omega=-[R/(c_p p)]Q(p) \quad , \tag{1d}$$

$$\partial_t M+(1/g)\int_0^{p_s} \nabla\cdot(\bar{q}(p)\bar{v})\,dp=E_v-P_r \quad , \tag{1e}$$

where M is the total moisture per unit area, E_v and p_r are evaporation and precipitation rate, respectively. The mean basic state is assumed to be motionless with mean specific humidity (wang, 1988)

$$\bar{q}(p)=\bar{q}_s(p/p_s)^{H/H1-1} \tag{2}$$

where H and H_1 are density scale height and water vapor scale height, respectively. The mean surface specific humidity is assumed to be a function of sea surface temperature (Wang, 1988),

$$\bar{q}_s=(0.94SST(^oc)-7.46)\times10^{-3} \quad , \tag{3}$$

while the SST is assumed to be a function of latitude only:

$$SST(y)=SSTM\ \exp(-\beta y^2/15c_o)-\beta y^2/c_o \quad . \tag{4}$$

In equations (1a-e), the frictional forces F_x, F_y are taken into account only in the boundary layer. To focus on anisotropic planetary scale motion, we will adopt semi-geostrophic approximation by neglecting $\partial v/\partial t$ in Eq. (1b).

The diabatic heating rate Q includes only longwave radiation cooling $Q_1=\mu(c_p p/R)\partial\phi/\partial p$, and condensational heating Q_2 which is related to precipitation rate P_r by

$$\int_0^{p_s} Q_2 \frac{dp}{g} = L_c P_r \quad . \tag{5}$$

A crucial assumption for moisture conservation equation is that the precipitation rate is linearly balanced by the moisture convergence. This leads to

$$\int_0^{p_s} Q_2(p)\,dp = L_c \int_0^{p_s} \nabla\cdot(\overline{q\mathbf{v}})\,dp \quad .$$

3. MODEL EQUATIONS

Consider a simplified 2-1/2 layer model which consists of a two-level representation of free atmosphere and a well-mixed boundary layer (Wang, 1988). If we define barotropic and baroclinic parts of wind and geopotential by

$$u_\pm=(1/2)(u_3\pm u_1) \;,\quad v_\pm=(1/2)(v_3\pm v_1) \;,\quad \phi_\pm=(1/2)(\phi_3\pm\phi_1) \;, \tag{6}$$

the nondimensional governing equations for 2-level free atmosphere can be written as

$$\partial_t u_- - y v_- = -\partial_x \phi_- \quad , \tag{7a}$$

$$y u_- = -\partial_y \phi_- \quad , \tag{7b}$$

$$(\partial_t + N)\phi_- + (1-I)(\partial_x u_- + \partial_y v_-) = \omega_e(B-1) + \omega_u(I-1) \quad . \tag{7c}$$

$$\partial_t u_+ - y v_+ = -\partial_x \phi_+ \quad , \tag{7d}$$

$$y u_+ = -\partial_y \phi_+ \quad , \tag{7e}$$

$$\omega_e - \omega_u = -(\partial_x u_+ + \partial_y v_+) \quad , \tag{7f}$$

In (7), ω_u and ω_e are the vertical p-velocity at upper boundary p_u and the top of the boundary layer p_e, respectively; the nondimensional numbers are

$$N=\mu/\sqrt{\beta c_0} \quad , \tag{8a}$$

$$I=L_c(\bar{q}_s-\bar{q}_1)/S \quad , \tag{8b}$$

$$B=L_c(2\bar{q}_e-\bar{q}_3-\bar{q}_1)/S \quad , \tag{8c}$$

$$S=(c_p p_2 \Delta p/R)S_2 \quad , \tag{8d}$$

where N is nondimensional Newtonian cooling coefficient, S measures mean static stability of the basic state; I and B represent ratios of latent heating to adiabatic cooling due to vertical motion at p_2 and p_e, respectively. We have used c_0, $(c_0/\beta)^{1/2}$, $(\beta c_0)^{-1/2}$, c_0^2, and $2\Delta p(\beta c_0)^{1/2}$ as horizontal velocity, length, time, geopotential, and vertical p-velocity scales, respectively, where c_0 is long gravity wave speed in 2-level model.

The boundary layer friction-induced vertical velocity ω_e can be expressed in terms of geopotential at p_e and ϕ_e (Wang and Chen, 1988):

$$\omega_e=\frac{(p_s-p_e)}{2\Delta p(E^2+y^2)}\left[\frac{2y}{E^2+y^2}(E\partial_y-y\partial_x)-E(\partial^2_{xx}+\partial^2_{yy})+\partial_x\right]\phi_e\ , \qquad (9)$$

where $E=\rho_e gA_z/[(\rho_s-\rho_e)\sqrt{\beta c_0}\cdot h\cdot \ln(h/z_0)]$ is the Ekman number determined by turbulent viscosity A_z, depth of the surface layer h, the surface roughness length z_0 and density ρ_e.

The upper boundary and lateral boundary conditions are

$$\omega_u=0\ , \quad \text{at } p=p_u\ , \qquad (10a)$$

$$v_+=v_-=0\ , \quad \text{at } y=\pm y_0\ , \qquad (10b)$$

Normal mode solutions were searched for using both shooting method and finite difference method. the stability of the moist atmosphere to semi-geostrophic perturbations is analyzed and the results are summarized in the next three sections.

4. MODE SELECTION

Fig.1 shows the phase speed and growth rate as function of SST for Kelvin and m=1,2 Rossby waves, where m is the meridional index. The moisture content increases with increasing SST. Without sufficient moisture supply, the model atmosphere is stable. Due to the presence of thermal and boundary layer dissipations, westward moving Rossby waves decay faster than eastward moving Kelvin waves. When

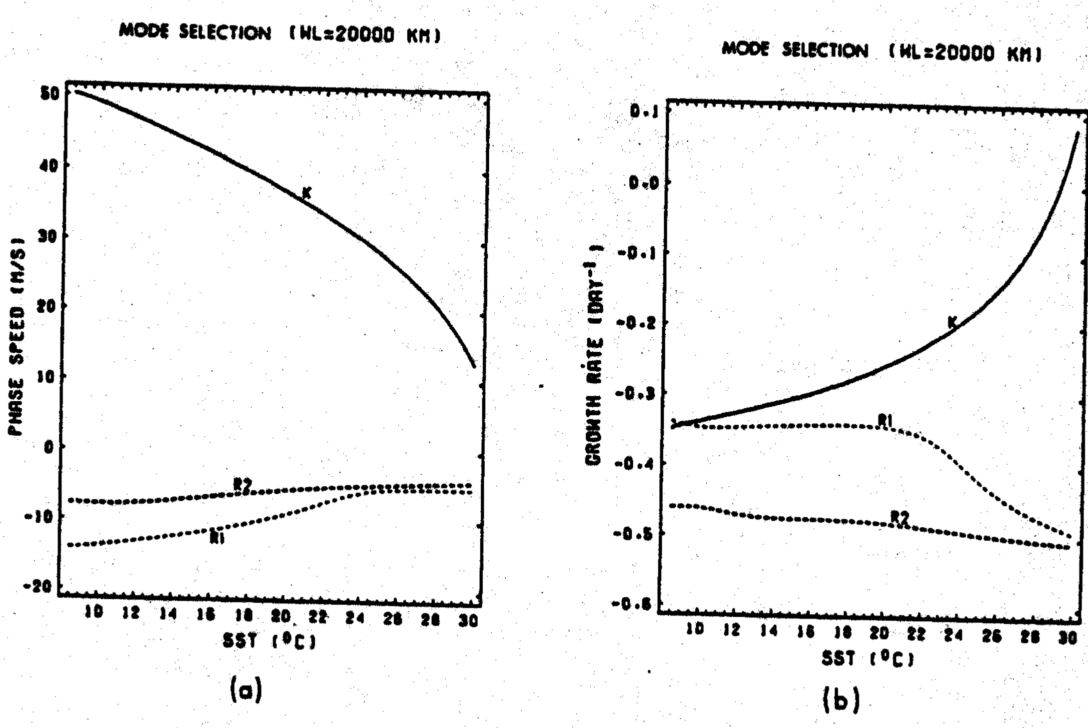


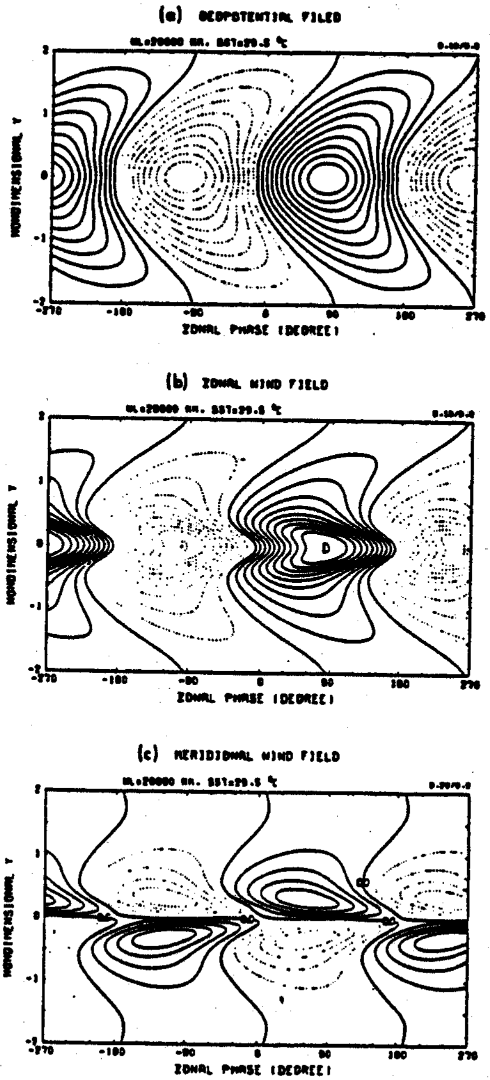
Fig. 1. (a) Phase speed and (b) growth rate of the moist Kelvin mode (K), m=1 Rossby mode (R1), and m=2 Rossby mode (R2) as functions of SSTM. The wavelength is 20,000 km.

moisture concentration gradually increases, the moist Kelvin mode becomes progressively less damped and finally begins to grow when SST exceeds an critical value. On the other hand, moist Rossby waves are always damped. The unstable mode selection in the present model can be explained in terms of wave energy generation due to latent heating induced by the boundary layer frictional moisture convergence. Because of the large concentration of moisture in the boundary layer, the latent heating associated with frictional moisture convergence produces a substantial portion (about 1/3) of the wave energy. Since the availability of basic-state moist static energy is highest at equator, the equatorial warm region is most conducive (or destructive) for wave energy generation. In this region, both Kelvin and Rossby wave-induced boundary layer convergences reach their maxima (Figures not show); however, the frictional upward motion is positively correlated with temperature in the Kelvin mode, while the negative covariance between them is found for the Rossby mode. Thus, available potential energy is generated efficiently by the frictional convergence-induced latent heating in the moist Kelvin mode, but is destroyed in the moist Rossby modes.

5. THE EFFECTS OF THE HORIZONTAL MODE COUPLING

Although unstable modes appear to be rooted in Kelvin waves, they are modified by the dynamic coupling with Rossby waves. The horizontal structures of the zonal wind and geopotential of the unstable mode shown in Fig.2 resemble those of Kelvin waves but exhibit significant meridional wind components, which are asymmetric about the equator and similar to those of the lowest meridional mode of Rossby wave. This suggests that the unstable mode is in nature a moist Kelvin mode modified through coupling with a long Rossby wave.

The coupling between moist Kelvin and Rossby modes via



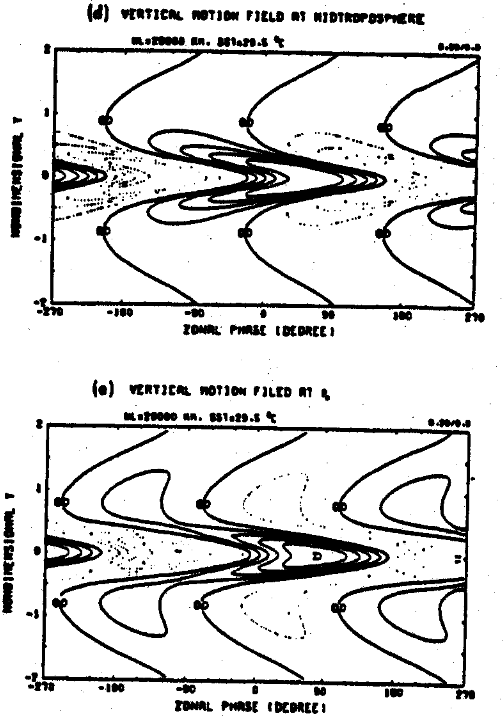


Fig. 2. Horizontal structure of the unstable wavenumber two at SSTM=29.5°C: (a) geopotential ϕ_-, (b) zonal wind, u_-, (c) meridional wind, v_-, (d) ω_2, and (e) ω_e. Solid and dotted lines denote positive (or zero) and negative contours, respectively. Contour intervals for ϕ_- and u_- are 10% of the maximum value and those for v_-, ω_2, and ω_e are 20% of the maximum value.

boundary layer frictional convergence and associated latent heating has fundamental impacts on wave instability. Fig. 3 compares growth rates and phase speeds computed from non-coupled moist Kelvin model (dotted) and moist coupled Kelvin-Rossby models under constant SST (dashed-dotted) and a latetude-dependent SST (solid). It is seen that the horizontal mode-coupling acts as an efficient brake to reduce the eastward propagation speed (Fig.3b), suppresses unrealistically fast growth of the uncoupled moist Kelvin mode by creating significant meridional flows (Fig.3a). More importantly it favors the amplification of long planetary waves, rather than short waves, providing a longwave selection mechanism (Fig. 3a).

6. THE SEASONALITY OF THE INTRASEASONAL OSCILLATION

By analyzing 19 near equatorial station rawinsonde data, Madden (1986) showed that the intraseasonal variability in zonal wind exceeds that in adjacent lower and higher frequency bands by the largest amount during December, January, and February. Using ten years of outgoing longwave radiation (OLR) data, we investigated temporal variations of 77 low frequency equatorial eastward-moving events and found that the overall intensity of intraseasonal convective anomalies are significantly stronger in boreal winter (from November to April) than in boreal summer (Wang and Rui, 1989).

The annual variation of equatorial intraseasonal wave activity may be caused by the annual march of the "thermal" equator where the highest SST is located. This notion is confirmed by an experiment in which the maximum SST is shifted to 7.5^{o} N, a situation occurring during boreal summer. In this case, the growth rate of the unstable coupled Kelvin-Rossby mode is substantially reduced (Fig.4). The separation of the thermal effect of warm ocean water

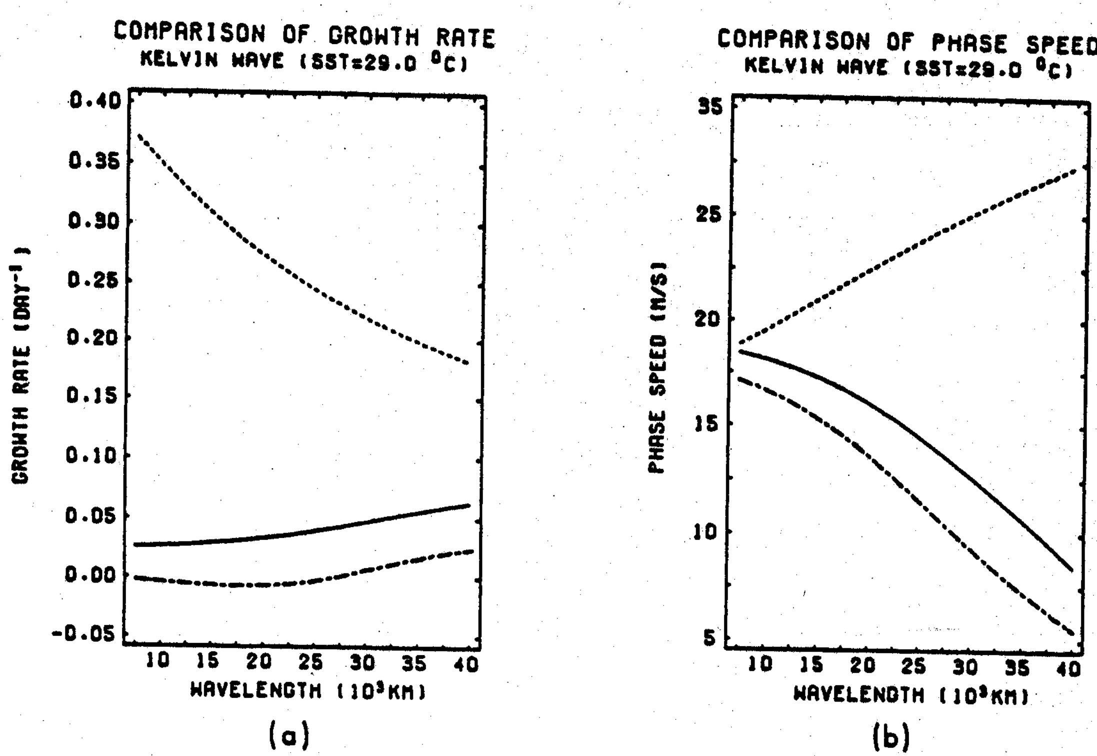


Fig. 3. Comparison of (a) growth rate and (b) phase speed as functions of wavelength. Dotted, dash-dotted and solid lines indicate values computed from the viscous Kelvin wave-CISK model, the model with uniform SST=29°C, and the model with latitude dependent SST, (4) , respectively. All parameters are the same for three cases.

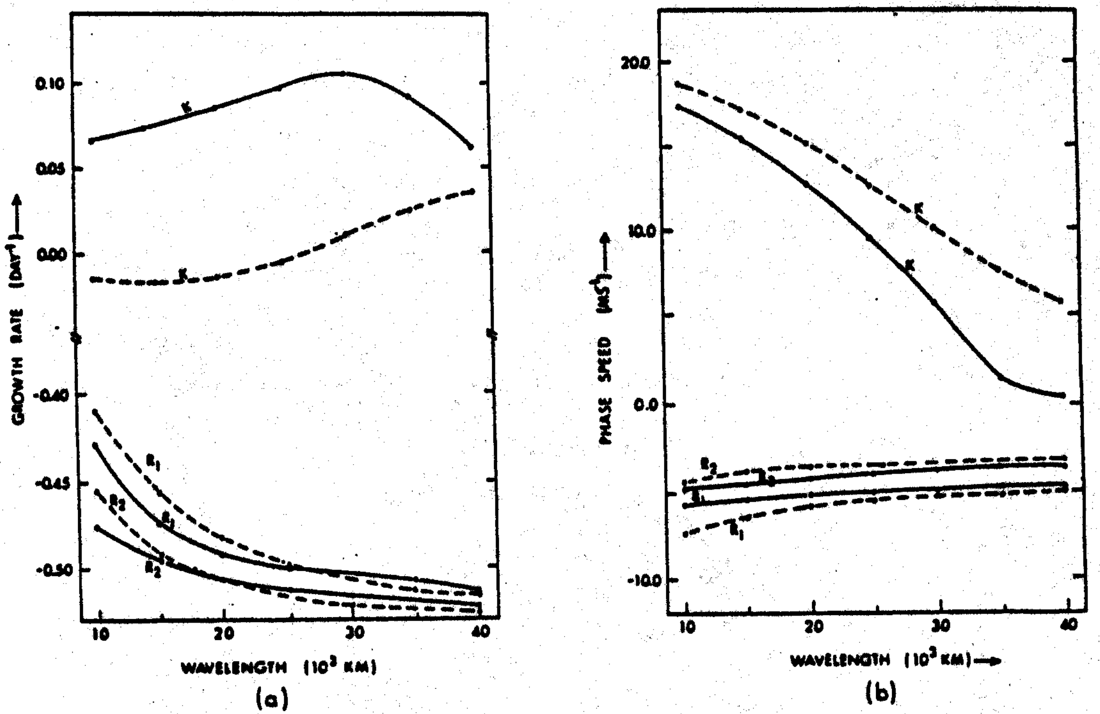


Fig. 4. Comparison of (a) growth rate and (b) zonal phase speed of the unstable mode computed using SST profile with maximum at the equator (solid lines) and a profile with maximum at 7.5°N (dashed lines). In both cases the maximum SST is 30°C.

from the dynamic effect of the equator stabilizes the tropical atmosphere for planetary scale low frequency disturbances.

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