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THE ROLE OF MERIDIONAL WIND STRESS IN THE TROPICAL UNSTABLE AIR-SEA INTERACTION

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ABSTRACT: With a simple tropical coupled ocean-atmosphere model, this paper presents an analysis aiming to understand the relative role of the meridional and zonal wind stresses in the tropical unstable air-sea interaction. The roles of the zonal wind stress, the meridional wind stress and the both are considered respectively into the coupled system. It is demonstrated that the meridional component of the wind stress does not lead to any instability under the local thermal balance assumption, but it does lead to a weak instability under the sea surface temperature advection assumption. Unstable air-sea interaction is dominated by the zonal component of the wind stress, suggesting that ignoring the meridional wind stress is approximately feasible in studying the tropical unstable air-sea interaction.

Key words: wind stress; air-sea interactions; instability

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1 INTRODUCTION

As shown in observational facts, the evolution of El Niño displays as the amplitude expansion and migration of anomalous disturbances in air-sea coupling systems. Theoretical address of the phenomenon has been attempted from the aspect of coupling dynamics. The results show that the air-sea interaction system includes a kind of unstable mode, which is propagating in directions determined by factors governing SST changes.

With an air-sea coupling model, Lau[1] was among the earliest people who discovered unstable stationary wave modes in the air-sea interactions. Using the traditional local thermodynamic equilibrium assumption, in which SST is proportional to the thickness of the thermocline, Philander et al. [2] decided that the unstable modes of the interaction were slowly propagating eastward in the oceanic Kelvin waves. Assuming that SST was determined by advections, Rennick[3] and Gill[4] had the Rossby solutions to unstable weatward propagation. Hirst[5] argued that there was some kind of unstable east-traveling mode between the results of Philander and Gill if SST is set to relate to advection as well as the thermocline thickness. Chao and Zhang[6] discussed the issue of unstable air-sea interactions for the viewpoint of waving and pointed out that the coupled waves can also move east and generate instability via the interaction even if the Kelvin wave is removed from the model. In their study of the coupling nature for various kinds of waving in the ocean and atmosphere, Yang et al[7, 8]. reported loss of stability in oceanic Kelvin waves due to air-sea interactions having an advantageous zonal scale of ~ 10^4 km in the tropics, which was the main result of coupling between the atmospheric Rossby wave and oceanic Kelvin wave. It is then seen that the wave action in the air-sea interaction and its instability are unique and they are accountable not only by dynamics of single components of the ocean or atmosphere but also resulted from air-sea interactions. The theory of unstable air-sea

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interactions generally justifies the large-scale air-sea interactions as determined by Bjerknes[9] and explaines the mechanisms with which the ENSO episodes evolve.

In previous work, however, much emphasis has been on the role of zonal wind stress in the air-sea coupling while leaving the role of meridional wind stress much ignored or neglected. The question whether the latter has any intrinsic effect on unstable air-sea interactions remains unanswered and there has not been any sound theoretic foundations for its negligence.

Addressing the issue, the current work employs a simple air-sea coupled tropical model to study the relative roles of meridional and zonal wind stress in the formation of unstable air-sea coupled modes, with the focus on the former.

2 AIR-SEA COUPLED EQUATION SETS AND SIMPLIFICATION

The atmospheric components in the coupled model are the first tropical baroclinic modes that are linear and approximated under the longwave, which are expressed on the equatorial $\boldsymbol{\beta}$ plane as:

$$\frac{\partial u_a}{\partial t} - byv_a = -\frac{\partial j_a}{\partial x} \tag{1}$$

$$byu_a = -\frac{\partial j_a}{\partial y} \tag{2}$$

$$\frac{\partial j_a}{\partial t} + c_a^2\left(\frac{\partial u_a}{\partial x} + \frac{\partial v_a}{\partial y}\right) = -Q \tag{3}$$

in which the subscript "*a*" stands for the atmosphere, u_a, v_a and j_a for the perturbed wind speed in the meridional and zonal directions and atmospheric pressure, c_a for phase velocity of the gravity wave in non-rotational atmosphere and Q for diabatic heating rate that links with SST.

The oceanic components in the coupled model are a set of shallow motion equations that are linear with reduced gravity, which are expressed on the equatorial $\boldsymbol{\beta}$ plane as:

$$\frac{\partial u}{\partial t} - byv = -g'\frac{\partial h}{\partial x} + t_x \tag{4}$$

$$byu = -g'\frac{\partial h}{\partial y} + t_y \tag{5}$$

$$\frac{\partial h}{\partial t} + H_0\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0 \tag{6}$$

in which u and v are for the zonal and meridional velocity of perturbed ocean currents, h is the disturbance in the depth of the mixed oceanic layer, g' is the reduced gravity acceleration, H_0 is the mean depth of the mixed layer, t_x is the zonal wind stress and t_y is the meridional wind stress.

For the oceanic components in the couple model, the thermodynamic equation describes the changes in SST, which can be simply put as

$$e\frac{\partial T}{\partial t} + Au - Kh + a_s T = 0 \tag{7}$$

in which T is the disturbance of SST, A is the zonal gradient of the climatologically averaged SST, K is a constant associating with climatologically averaged up-welling currents, a_s is the Newtonian cooling coefficient. For e, the value takes 0 and 1, depicting two typical cases of local SST changes: When $\varepsilon=0$, take $A=0$, then Eq.(7) is changed to $T = Kh/\alpha_s$, meaning that SST anomalies are proportional to the disturbance of the mixed layer thickness, which is a traditional assumption for local thermodynamic equilibrium; when $\varepsilon=1$, take $K=0$, $\alpha_s = 0$, then we have $\partial T/\partial t + Au = 0$, i.e. SST changes are governed by advections only, which is the so-called advection assumption.

To enclose the equations above, the air-sea coupling relations must be determined first. The anomalies of SST can cause anomalies in atmospheric evaporation and latent heat release by condensation, which in turn heat the atmosphere and results in wind field anomalies; the wind field anomalies in turn drive ocean current anomalies by way of wind stress. Such process of coupling dynamics and thermodynamics can be linearly expressed as

$$Q = c_a \alpha T, \quad \tau_x = \gamma u_a, \quad \tau_y = \gamma v_a \tag{8}$$

in which α and γ are respectively the coefficients for thermodynamic and dynamic coupling between the air and sea. The values take $\gamma = 5.0\times10^{-7}\ \mathrm{s}^{-1}$ and $\alpha = 6.75\times10^{-5}\ \mathrm{s}^{-1}$, respectively.

To solve the above equation sets, the Gill transform is introduced to the atmosphere and ocean[10], in other words,

$$q_a = \frac{\varphi_a}{c_a} + u_a, \qquad r_a = \frac{\varphi_a}{c_a} - u_a \tag{9}$$

$$q = \frac{g'h}{c_0} + u, \qquad r = \frac{g'h}{c_0} - u \tag{10}$$

are defined. Here, $c_0 = \sqrt{g'H_0}$ is the characteristic velocity of the ocean. When $a_a = \sqrt{c_a/2\beta}$ and $a_0 = \sqrt{c_0/2\beta}$ are defined to be the Rossby's deformed radii for the atmosphere and ocean, respectively, with $Y_a = y/a_a$, $Y_0 = y/a_0$ being assumed, then in relation to the dimensionless coordinates Y_a, we have in the y direction a parabolic cylinder function expansion of various variables in the atmospheric equation sets as in

$$(q_a, r_a, v_a, \frac{Q}{c_a}) = \sum_{n=0}^{\infty} (q_{an}, r_{an}, v_{an}, F_{an}) D_n(Y_a) \tag{11}$$

and in relation to the dimensionless coordinates Y_0, we have in the y direction a parabolic cylinder function expansion of various variables in the atmospheric equation sets as in

$$(q, r, v, \tau_x, \tau_y, T) = \sum_{n=0}^{\infty} (q_n, r_n, v_n, F_{xn}, F_{yn}, T_n) D_n(Y_0) \tag{12}$$

Firstly, Eqs.(9) and (10) are used to transform Eqs.(1) – (8). Then, Eqs.(11) and (12) are introduced to the equation set for the air-sea coupling that has been transformed to derive an expanded set of atmospheric and oceanic states (equations omitted), on the basis of the orthogonality of the parabolic cylindrical function. Specifically, from Eq.(8), the air-sea coupling relation, or terms of interaction, F_{an}, F_{xn} and F_{yn}, can be determined with the following expressions

$$F_{an} = \frac{1}{n!\sqrt{2\pi}} \int_{-\infty}^{\infty} \frac{Q}{c_a} D_n(Y_a)\,\mathrm{d}Y_a = \frac{1}{n!\sqrt{2\pi}} \int_{-\infty}^{\infty} \alpha T D_n(Y_a)\,\mathrm{d}Y_a \tag{13}$$

$$F_{xn} = \frac{1}{n!\sqrt{2\pi}} \int_{-\infty}^{\infty} \tau_x D_n(Y_0)\,\mathrm{d}Y_0 = \frac{1}{n!\sqrt{2\pi}} \int_{-\infty}^{\infty} \frac{1}{2}\gamma (q_a - r_a) D_n(Y_0)\,\mathrm{d}Y_0 \tag{14}$$

$$F_{yn} = \frac{1}{n!\sqrt{2p}}\int_{-\infty}^{\infty} t_y D_n(Y_0)\,\mathrm{d}Y_0 = \frac{g}{n!\sqrt{2p}}\int_{-\infty}^{\infty} v_a D_n(Y_0)\,\mathrm{d}Y_0 \tag{15}$$

To have a quantitative discussion of the dynamics of the air-sea coupling, especially the wave interactions with varying atmospheric and oceanic nature, the expanded equation set needs to be reduced. The current reduction is carried out so as to cover basic nature of the physics: (1) In our dynamic equation set, the parabolic cylindrical function expansion is truncated where n = 1(i.e. by taking 0 and 1 in n) and the Yanai mode is omitted, with only the Kelvin wave and lowest-order Rossby wave retained for either the atmosphere or the ocean. (2) For the SST equation, the distribution of SST disturbance across the equatorial Pacific region is used to perform zero-order approximation on T_n (corresponding to the fact that the maximum of SST disturbance is on the equator and symmetric about it). (3) In the equations, the atmosphere is viewed as quasi-stationary response, i.e. the atmosphere has a rapid adaptation to the variation of the ocean within the time scale of the latter. It is then possible to set $\partial/\partial t = 0$ in the atmospheric equations. Following the approximations, a reduced set of air-sea coupled equations can be derived as in

$$c_a \frac{\partial q_{a0}}{\partial x} = -a\, l_1 T_0 \tag{16}$$

$$-\frac{1}{3} c_a \frac{\partial q_{a2}}{\partial x} = -a\, l_2 T_0 \tag{17}$$

$$\frac{\partial q_0}{\partial t} + c_0 \frac{\partial q_0}{\partial x} = g\,(l_3 q_{a0} - l_4 q_{a2}) \tag{18}$$

$$\frac{\partial q_2}{\partial t} - \frac{1}{3} c_0 \frac{\partial q_2}{\partial x} = g\,(-l_5 q_{a0} + l_6 q_{a2}) + ag\, l_7 (\frac{\partial T_0}{\partial t} - c_0 \frac{\partial T_0}{\partial x}) \tag{19}$$

$$e \frac{\partial T_0}{\partial t} + (a_s + 3A_2 ag\, l_7) T_0 = 2A_2 q_2 - A_1 q_0 \tag{20}$$

Here, $S = (c_0 / c_a)^{1/2}$ is defined with l_i (i = 1, 7) being all positive constants related with S, and both $A_1 = (A - Kc_0 / g')/2$ and $A_2 = (A + Kc_0 / g')/2$ constants. Up till now, Eqs.(16) – (20) have constituted the initial set of equations for the questions concerned. What is being described by the equation is a process in which the atmospheric Kelvin wave (q_{a0}), the lowest-order Rossby wave (q_{a2}), the oceanic Kelvin wave (q_0) and lowest-order Rossby wave (q_2) interact among themselves via SST (T_0) and wind stress.

3 RELATIVE CONTRIBUTION OF MERIDIONAL AND ZONAL WIND STRESS TO UNSTABLE MODES

3.1 *Case with only longitudinal wind stress*

In the equation set for the ocean, set $t_x = 0$ and $t_y \neq 0$, then Eq.(16) – (20) is reduced to an air-sea coupling equation set under the effect of longitudinal wind stress

$$\begin{cases} c_a \dfrac{\partial q_{a0}}{\partial x} = -a l_1 T_0 \\ -\dfrac{1}{3} c_a \dfrac{\partial q_{a2}}{\partial x} = -a l_2 T_0 \\ \dfrac{\partial q_0}{\partial t} + c_0 \dfrac{\partial q_0}{\partial x} = 0 \\ \dfrac{\partial q_2}{\partial t} - \dfrac{1}{3} c_0 \dfrac{\partial q_2}{\partial x} = agl_7 (\dfrac{\partial T_0}{\partial t} - c_0 \dfrac{\partial T_0}{\partial x}) \\ e \dfrac{\partial T_0}{\partial t} + (a_s + 3A_2 agl_7) T_0 = 2A_2 q_2 - A_1 q_0 \end{cases} \tag{21}$$

A dispersion relation expression can be obtained by setting formal solution ($\propto e^{i(kx-wt)}$) that is substituted into Eq.(21). The solution to the third expression is a free Kelvin wave not being involved with air-sea coupling. The nature of the characteristic air-sea coupling modes is addressed in three cases.

For the first case, with the assumption of local thermodynamic equilibrium, i.e. taking $e=0$ and $A = 0$, then $A_1 = -Kc_0 / 2g'$, $A_2 = Kc_0 / 2g'$ and we have the solution to Eq.(21) as in

$$w = -\frac{1}{3} c_0 k (1 - \frac{4A_2 agl_7}{a_s + A_2 agl_7}) \tag{22}$$

It is the Rossby solution subject to air-sea interactions. Fig.1c gives its dispersion relation and the distribution of growth rate with the wavenumber. It is shown that with the assumption of local thermodynamic equilibrium, the air-sea interaction is weak if it is powered by longitudinal wind stress only, i.e. the value within the brackets of Eq.(22) approaches to unity, so that the coupled wave is made very close to the free Rossby wave of the ocean. It is also known from Eq.(22) that there is no imaginary part in the dispersion relation expression for Rossby wave dispersion so that no unstable growth will occur.

For the second case, with the assumption of zonal temperature advection, i.e. with $e=1, a_s=0, K=0$, we have $A_1 = A_2 = A/2$. In this circumstance, there are two solutions to the dispersion relation derived from Eq.(21), which are given in Fig.2c. The two solutions can be rewritten approximately as follows:

$$w_+ \approx \frac{3}{2} Aagl_7 i \tag{23}$$

$$w_- \approx -\frac{1}{3} c_0 k - 2Aagl_7 i \tag{24}$$

in which w_+ is for the mode of temperature advection with the real part approximately being zero. Its velocity of transportation is also zero in approximation and generally stationary. In the meantime, the mode is decaying at a rate being constant in approximation as a、g、l_7>0 and A<0 for the equatorial Pacific Ocean. w_- is for the mode of the Rossby wave which is subject to air-sea interactions with the propagation speed approximately the same as the free Rossby wave, though weak instability is found with w_- over the entire wavelength and the growth rate is also approximately constant for the instability. Fig.2c gives the variation with wavenumber of the frequency and growth rate for the two solutions, with the characteristics being consistent with solutions approximately determined.

For the third case, with the assumption that SST is both related with the thickness of the mixed layer and with the advection process, i.e. setting $e=1$ only, the two solutions to the dispersion relation derive from Eq.(21) can be determined, which are given in Fig.3. The solutions are respectively for temperature advection mode and the oceanic Rossby wave subject to the air-sea interaction and their dispersion relation, phase velocity and distribution of the growth rate with the wavenumber are presented in Fig.3. It is seen that it is a result between local thermodynamic equilibrium assumption and temperature advection assumption. The oceanic Rossby wave is similar to that under the assumption of temperature advection, which grows at a constant but small rate over the entire wavelength; the advection mode can be considered stationary and fast-decaying.

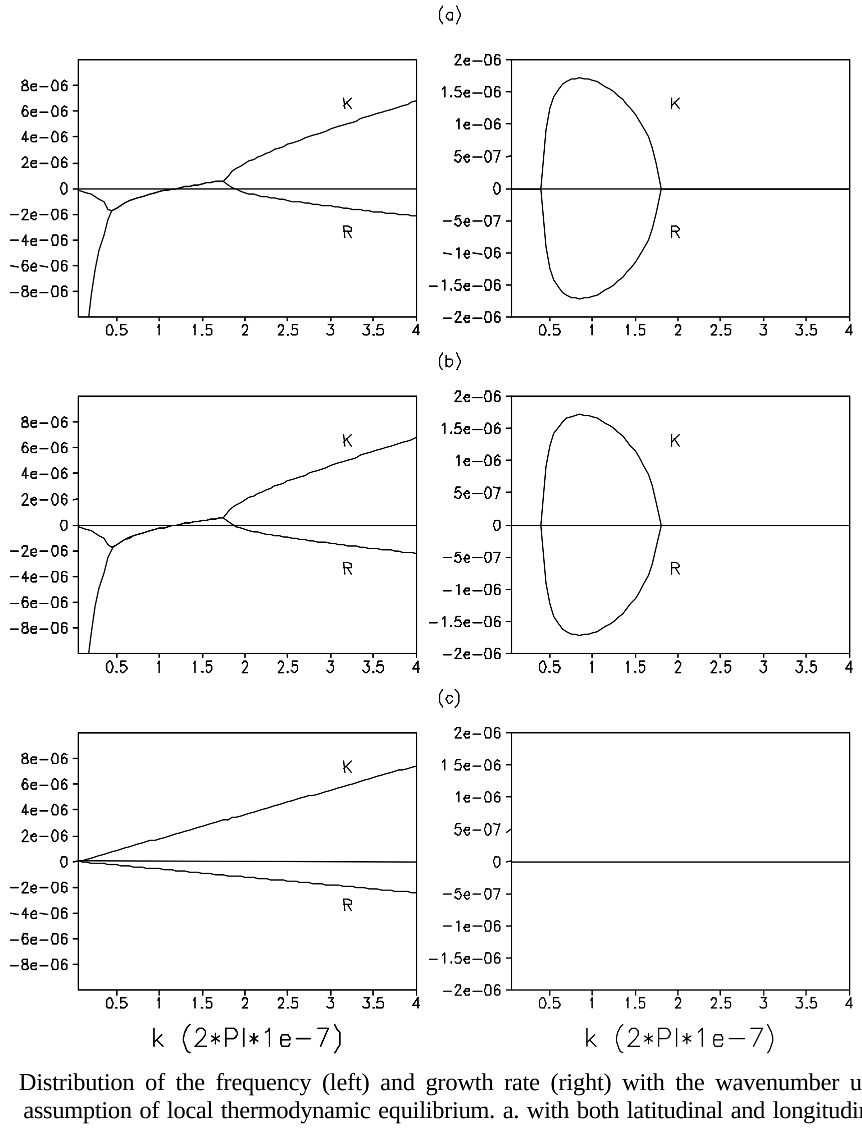

Fig.1 Distribution of the frequency (left) and growth rate (right) with the wavenumber under the assumption of local thermodynamic equilibrium. a. with both latitudinal and longitudinal wind stress; b. with latitudinal wind stress only; c. with longitudinal wind stress only. “K” and “R” are for Kelvin wave and Rossby wave, respectively.

3.2 *Case with only latitudinal wind stress*

In the equation set for the ocean, set $t_y = 0$ and $t_x \neq 0$, then Eq.(16) – (20) is reduced to an air-sea coupling equation set under the effect of latitudinal wind stress.

$$\begin{cases} c_a \dfrac{\partial q_{a0}}{\partial x} = -a\,l_1 T_0 \\ -\dfrac{1}{3} c_a \dfrac{\partial q_{a2}}{\partial x} = -a\,l_2 T_0 \\ \dfrac{\partial q_0}{\partial t} + c_0 \dfrac{\partial q_0}{\partial x} = g\,(l_3 q_{a0} - l_4 q_{a2}) \\ \dfrac{\partial q_2}{\partial t} - \dfrac{1}{3} c_0 \dfrac{\partial q_2}{\partial x} = g\,(-l_5 q_{a0} + l_6 q_{a2}) \\ e\dfrac{\partial T_0}{\partial t} + a_s T_0 = 2A_2 q_2 - A_1 q_0 \end{cases} \tag{25}$$

Likewise, formal solutions can be set to substitute into the equation set Eq.(25) for the expression of dispersion relation. The nature of its air-sea coupling modes will be discussed in two cases.

Firstly, there are two solutions to the dispersion relation expression for Eq.(25) under the assumption of local thermodynamic equilibrium. Fig.1b gives the dispersion relation of the solutions (in the left panel) and the variation of the coupled wave growth rate with the wavenumber (in the right panel). It shows that two kinds of waves are generated from the air-sea interaction with the assumption; the first kind of wave (w_+) is close to the oceanic free Kelvin wave (with a phase velocity of $c_0 = 3$ m/s) and the second kind of wave (w_-) is close to the oceanic free Rossby wave (with a phase velocity of $-c_0/3 = -1$ m/s). In essence, the two waves reflect the oceanic Kelvin wave and Rossby wave being subject to air-sea interactions. With the increase of wavelengths, the eastward propagation speed of the Kelvin wave decreases while the Rossby wave changes from westward propagation to eastward propagation. When the wavelength grows to 5700 km and beyond, the Kelvin wave starts to show instability while the Rossby wave starts decaying. Here, both of the waves have the same speed of propagation. The wavelength over which the Kelvin wave behaves unstably is between 5700 km and 20000 km and the rate of unstable growth has the maximum around 12000 km. When the wavelength is between 5700 km and 8000 km, the unstable Kelvin wave travels east and at a decreasing speed, with the increase of wavelength. It changes to a stationary wave when the wavelength is 8000 km and propagates, with increasing speed, instability to the west with the increase of wavelength, over the range between 8000 km and 20000 km. When the wavelength is larger than 20000 km, however, the instability is no longer present and the system shows the two kinds of wave again —the Kelvin wave slowly propagates west while the Rossby wave rapidly propagates west. It can be seen that under the assumption of local thermodynamic equilibrium, the air-sea interaction with the zonal wind stress only can cause the instability of the oceanic Kelvin wave, whose generation and propagation direction are linked with the zonal scale of the disturbance.

Secondly, with zonal temperature advection assumed, there are three solutions which is derived from Eq.(25), which is for the dispersion relation expression. Fig.2b gives the dispersion relation of the three waves (in the left panel) and their growth rates with wavenumber (in the right panel). It is seen from the left panel that as a reflection of the oceanic Kelvin wave and Rossby wave subject to the air-sea interaction, the first and second kinds of wave are very close to the free Kelvin wave and free Rossby wave for the ocean; the third kind of wave is a temperature advection mode caused by the advection assumption, which is approximately a stationary wave. From the variation of growth rates of the three kinds of wave with wavenumber (the right panel), we learn that (1) the Kelvin wave is decaying over the entire range of the wavelength; (2) the growth of the temperature advection mode is extremely weak (the rate being less than $10^{-10}\,\mathrm{s}^{-1}$) when the wavelength is larger than 80000 km while attenuation is the dominant fact throughout the rest of the wavelength; (3) though Rossby wave instability occurs over the entire wavelength, it is quite weak, with the unstable growth rate approaching zero when the wavelength is less than 10000 km, but rapidly increasing with the wavelength beyond the 10000 km mark —the longer the wavelength, the stronger the instability.

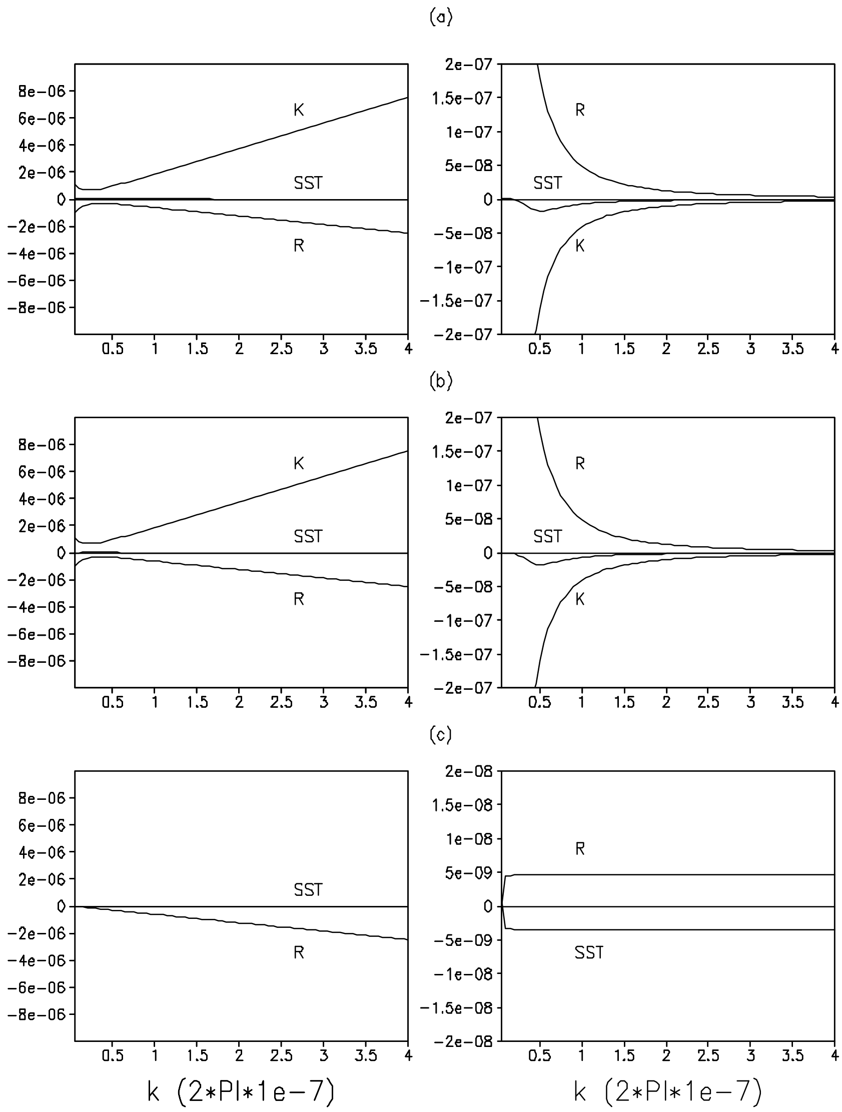

Fig.2 Distribution of the frequency (left) and growth rate (right) with the wavenumber under the assumption of zonal temperature advection. Legends for (a), (b) and (c) are the same as in Fig.1. “SST” is for the wave modes of SST.

3.3 *Case with both longitudinal and latitudinal wind stress*

Eqs.(16) – (20) have included the effect of longitudinal and latitudinal wind stress. The dispersion relation can be determined by substituting the formal solution into the equation set. Then, characteristic atmospheric and oceanic modes, determined by it, will be discussed.

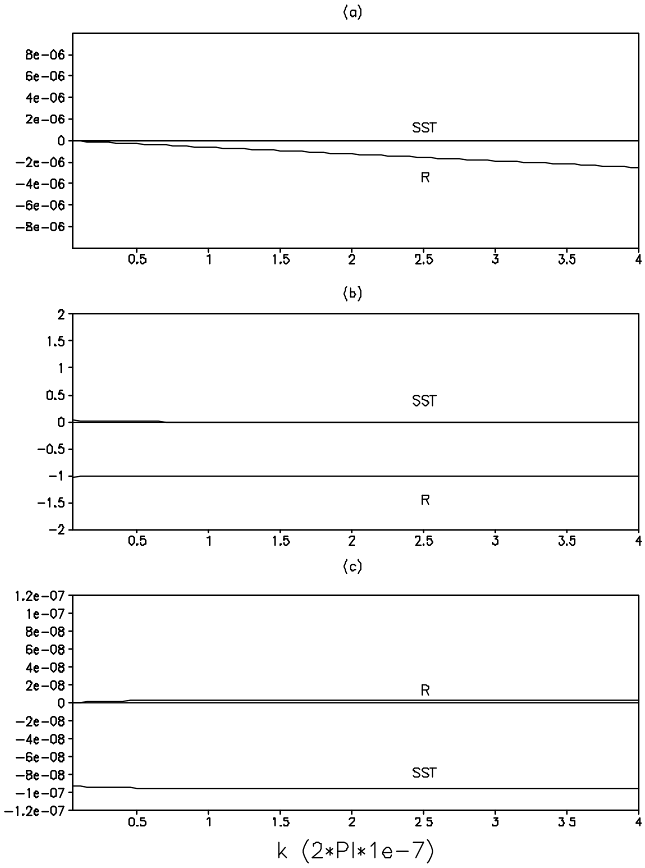

Fig.3 Distribution of the characteristic quantities of the coupled wave with the zonal wavenumber under the sole condition of longitudinal wind stress. a. frequency; b. phase speed; c. growth rate. "R" and "SST" stand for the Rossby mode and SST mode, respectively.

Firstly, with the assumption of local thermodynamic equilibrium, Fig.1a gives the dispersion relation for the two roots (as in the left panel) and the distribution of growth rate with the zonal wavenumber (as in the right panel). It is known by comparing Fig.1b with Fig.1c that the result as obtained by taking into account both the latitudinal and longitudinal wind stress (Fig.1a) is quite agreeable with that by considering the former stress only (Fig.1b). In other words, with the assumption of local thermodynamic equilibrium, the air-sea interaction gives rise to unstable oceanic Kelvin waves; the instability is totally contributed by the latitudinal wind stress as the single incorporation of the longitudinal wind effect does not lead to the appearance of instability, as suggested in the analysis above.

Secondly, with the assumption of zonal temperature advection, Fig.2a gives the dispersion relation for the three wave solutions (as in the left panel) and the variation of the growing rate with wavenumber (as in the right panel). At this stage, the air-sea interaction still leads to the instability of the oceanic Rossby wave. Comparing the characteristics in the three cases, we find that both longitudinal and latitudinal wind stress can cause the formation of the Rossby wave instability; as the unstable growth rate produced by the latitudinal wind is almost a magnitude of order larger than that by the longitudinal wind, the sum of total effect is highly similar to the result obtained with only the latitudinal wind stress considered.

4 CONCLUDING REMARKS

a. With the assumption of local thermodynamic equilibrium, the air-sea interaction causes unstable oceanic Kelvin wave. For this instability, the genesis and propagation directions are all determined by disturbance scale and no instability will be present unless the disturbance is within a particular range of wavelength (5700 km – 20000 km). The longitudinal wind stress alone will not cause unstable Kelvin wave and it can be stated that the unstable air-sea interaction with local thermodynamic equilibrium assumed is entirely caused by latitudinal wind stress.

b. With the assumption that SST is only determined by advection, the air-sea interaction results in the instability of the oceanic Rossby wave, which occurs over the whole wavelength. On the other hand, the instability is very weak, close to zero, when the wavelength is within 10000 km, but quite significant when it is larger than 10000 km and the longer the wavelength, the stronger the instability, which propagates westward. It is found from comparisons that both longitudinal and latitudinal wind stress can contribute to the instability of the oceanic Rossby wave, though the former has a very weak effect. It is therefore stated that the instability of the oceanic Rossby wave is also mainly contributed by the latitudinal wind stress.

c. Compared to the growth rate of instability of the oceanic Kelvin wave, the instability of the oceanic Rossby wave is weaker, with the range in which the maximum instability occurs moving in the direction of longwave. No matter with which instability, it is the latitudinal wind stress that is playing a leading role in air-sea coupling. Following what has achieved in the current work, we think that it is reasonable to ignore the role of the longitudinal wind stress in the study of the contribution of air-sea coupling dynamics in the tropical Pacific to the unstable evolution of ENSO.

It needs to be pointed out as a last point, however, that the current results are obtained with high-order truncation and more remains to be studied in the case of low-order truncation or more complicated coupling models..

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