ON THE INFLUENCES OF LARGE-SCALE INHOMOGENEITY OF SEA TEMPERATURE UPON THE OCEANIC WAVES IN THE TROPICAL REGIONS——PART I: LINEAR THEORETICAL ANALYSIS

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ABSTRACT

By using a linear oceanic mixed layer model, the influences of the horizontal gradients of sea surface temperature (SST) and the depth variations of the mixed layer upon tropical oceanic waves are investigated. The equatorial Rossby wave will be modified and a kind of slower thermal wave has been revealed under the influences of inhomogeneities of large-scale sea temperature field. An interesting result is that the propagating direction of the thermal wave is opposite to that of the classical Rossby wave. The result also shows that the thermal wave becomes dominant when the meridional gradient of sea temperature in the mixed layer exceeds a critical value. As a first approximation, it seems that both waves obtained by this study may be used to explain the observational facts that the SST anomalies can usually propagate in both directions, that is, eastward and westward, during the El Nino events.

I. INTRODUCTION

Observations have shown that the El Nino in the ocean and the southern oscillation in the atmosphere (ENSO) have important effects on the anomalous variations of both environments, particularly the global short-term variation of climate. The problem is arousing the interest of the academic circles of oceanography and meteorology in recent years[1-3].

On the aspect of the development of ENSO, an excellent review has been given by Philander[1]. In the typical ENSO, the westward trade-winds that prevail over most of the tropical Pacific Ocean converge on the low pressure zone where the air rises and where there is considerable cloudiness and rainfall. The air returns eastwards at greater altitudes and sinks over the cold, dry southeast high pressure zone at the eastern of the Pacific Ocean. This zonal cell is called the Walker Circulation. One of the precusors of El Nino is an eastward displacement of the upward branch of the Walker Circulation. The other precursor of ENSO is a southward displacement of ITCZ over the eastern Pacific. This displacement of the ITCZ is associated with light winds, unusually high SST and a deep thermocline in the southeastern equatorial Pacific Ocean. The second phase of the ENSO is the westward expansion of the anomalous condition that first appeared off the coast of Peru and Ecuador. The mature phase of ENSO is the unusually warm surface waters and exceptionally weak trade-winds over most of tropical Pacific Ocean. However, in the atypical 1982—1983 ENSO event, in early 1982 anomalous conditions in the western tropical Pacific Ocean were similar to the precusory ENSO conditions described above, namely an eastward displacement of the ascending branch of the Walker Circulation, the discrepancy is that the anomalous conditions continued to grow steadily in the western and central Pacific, so that there were eastward rather than westward winds over the western Pacific and abnormally high SST in the central Pacific until early autumn. Only then did the SST start to increase in the eastern tropical Pacific Ocean and subsequently spread westward. The main characteristic feature of the atypical ENSO is that the high SST did not appear off the coast of Peru and Ecuador at first.

It is comprehensible that the interaction between the ocean and atmosphere is of central importance during ENSO. Since the complexity of the air-sea interaction, in general, the characters of ENSO events are different from one another and it is even possible that different ENSO events may have different causes. At present, however, the investigations are focused on two aspects. One is where and when the onset of anomalous westerlies would take place, causing exceptional warm SST as the response of the ocean to the feeble winds. The other is the mechanism of the propagation or spread of the anomalously warm SST. A forcing theory for El Nino was usually attributed to Wyritki[4]. The equatorial easterlies over the central and/or western Pacific are anomalously strong, causing the sea level to rise in the west and fall in the east in the year preceding the El Nino event. Under favourable atmospheric circulations, the wind slackens, ultimately becoming westerly. Then the sea level reverses itself, falling in the west and rising in the east. And the water mass is transferred from west to east, with the thermocline layer thickened off the west coast of the South American Continent. The upwelling remains favorable, but the thermocline layer is now too deep to allow the cold water to reach the surface. The westward Rossby wave is activated by this propagation mentioned above in the east and the rising surface temperature along the coast is propagated by westward Rossby wave and/or spread by the other processes. However, this theory seems to be inconsistent. Two points, at least, may be discussed further. First, the 1982—1983 ENSO is a special event, the exceptional warm anomalous SST did not appear in the eastern Pacific at the beginning. Second, because the Kelvin wave can only exist on a line where the transverse velocities disappear, it is difficult to explain all the El Nino events with the help of Kelvin wave and Rossby wave.

At present, in order to illustrate some new observational facts, air-sea interaction theories are suggested[5,6], in which a theory named "Unstable air-sea interaction in the tropics" is an attractive one. However, the other possible mechanisms are also worthy of exploration. In our investigation, the influences of inhomogeneities of the large-scale SST upon the waves are studied. As well known, there is a cold water pool in the eastern Pacific Ocean (Fig. 1), so that the horizontal gradients of SST are different from the other regions. The horizontal distribution of SST and the depth of the mixed layer with annual variability are both caused by the large-scale and long-term air-sea interaction. Therefore, our investigation is a kind of air-sea interaction theories in a broad sense.

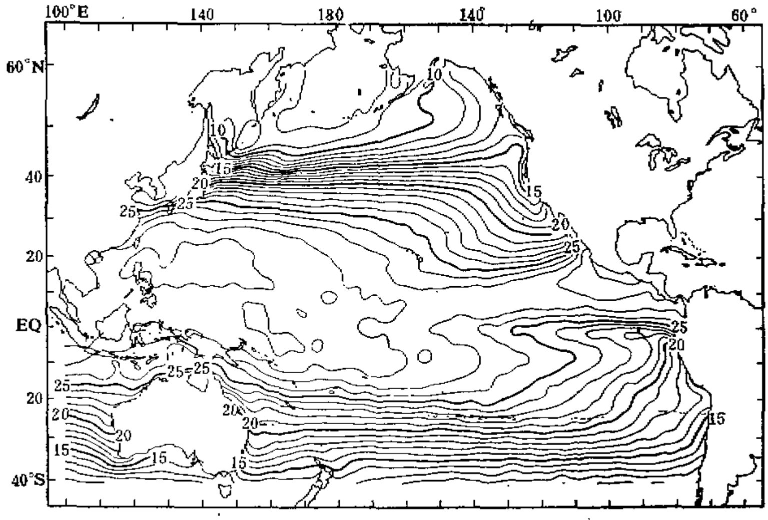


Fig. 1. Mean SST filtered climatology (in °C) on an one-degree grid for the Pacific Ocean for July[11].

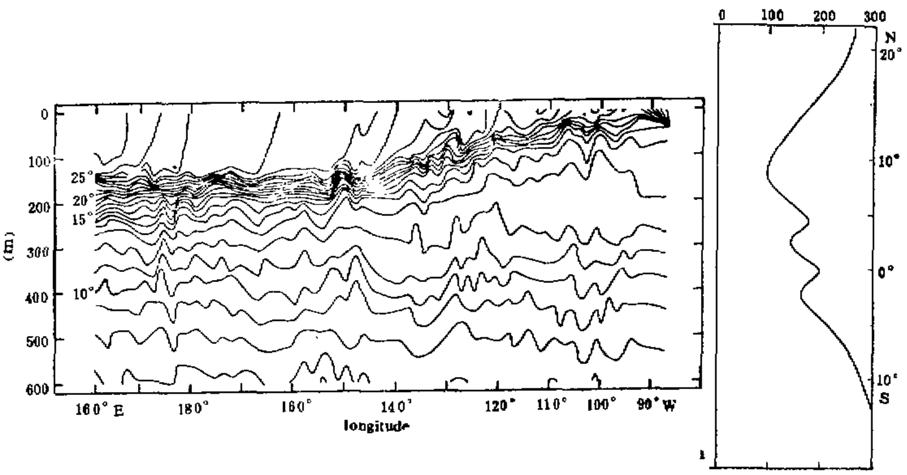


Fig. 2. Temperature as a function of longitude and depth along the equator in the Pacific Ocean as measured by Colin et al[12]. The panel on the right gives the mean profile of the 14°C isotherm as a function of latitude, showing the ridges and troughs[13].

II. MODEL

Assume $\bar{T}$ (x, y, z) to be the climatic large-scale sea temperature in the oceanic mixed layer. The large-scale climatic ocean currents adjusted to the sea temperature $\bar{T}$, strictly speaking, should be incorporated in the model. However, if only the vertical means of variables in the mixed layer are investigated, the contribution of large-scale oceanic currents will cancel out each other to a certain degree, because the currents at upper and lower layers, in general, are opposite in direction. For a preliminary study, the effects of climatic currents are neglected.

From Fig. 2, it can be seen that the vertical temperature distribution has quite a stable stratification, so that the heat and momentum exchange between the mixed layer and the thermocline is not important for the model's time scale of the order of one month. Thus, an equivalent shallow water model can be used in the mixed layer. The governing equation for the anomalous T_s' (perturbation temperature) with the linearized approximation is

$$\frac{\partial T_s'}{\partial t}+\frac{\Delta \bar{T}_z}{D}w_s'+u_s'\frac{\partial \bar{T}}{\partial x}+v_s'\frac{\partial \bar{T}}{\partial y}=0, \qquad (1)$$

where D is the depth of mixed layer and $\bar{T}_z=\bar{T}_{z=0}-\bar{T}_{z=-D}$, the difference of climatological sea temperature between the sea surface and the top of the thermocline. The symbol bar indicates the climatological value. In usual cases, heat is not full-mixed even in the mixed layer, $\Delta\bar{T}_z$ still has the value of the order of 10° °C. u_s', v_s' and w_s' are perturbations of the oceanic current in x, y and z directions respectively.

Integrating Eq. (1) from $-D$ to 0, the sea surface, we have

$$\frac{\partial \hat{T}_s'}{\partial t}+\frac{\Delta \bar{T}_z}{D}\hat{w}_s'+\hat{u}_s'\frac{\partial \bar{T}}{\partial x}+\hat{v}_s'\frac{\partial \bar{T}}{\partial y}=0 \quad , \qquad (2)$$

where

$$\hat{u}_s'=\frac{1}{D}\int_{-D}^{0}u_s'dz, \quad \hat{v}_s'=\frac{1}{D}\int_{-D}^{0}v_s'dz. \qquad (3)$$

Also, integrating the continuity equation

$$\frac{\partial u_s'}{\partial x}+\frac{\partial v_s'}{\partial y}+\frac{\partial w_s'}{\partial z}=0 \quad , \qquad (4)$$

with the help of the parameter variable integration and the boundary conditions

$$z=0,\ w_s'=0;\quad z=-D,\ w_s'\Big|_{z=-D}\doteq\hat{w}_s' \ , \qquad (5)$$

we have

$$\hat{w}_s'=D\left(\frac{\partial \hat{u}_s'}{\partial x}+\frac{\partial \hat{v}_s'}{\partial y}-\frac{\hat{u}_s'}{D}\frac{\partial D}{\partial x}-\frac{\hat{v}_s'}{D}\frac{\partial D}{\partial y}\right). \qquad (6)$$

Substituting Eq. (6) into Eq. (2), we finally obtain

$$\frac{\partial \hat{T}_s'}{\partial t}+\hat{u}_s'\left(\frac{\partial \bar{T}}{\partial x}-\frac{\partial \bar{T}_z}{D}\frac{\partial D}{\partial x}\right)+\hat{v}_s'\left(\frac{\partial \bar{T}}{\partial y}-\frac{\Delta \bar{T}_z}{D}\frac{\partial D}{\partial y}\right)$$

$$+\Delta\bar{T}_z\left(\frac{\partial \hat{u}_s'}{\partial x}+\frac{\partial \hat{v}_s'}{\partial y}\right)=0. \qquad (7)$$

On the other hand, if the frequencies of motion investigated here are much smaller than that of the inertial frequency, then the inertial motion with higher frequencies can be filtered out. In this case, the equations of motion without the effect of wind stress become[7]

$$f\hat{u}'_s = -\frac{1}{\rho_s}\frac{\partial \hat{p}'_s}{\partial y} - \frac{\beta}{f\rho_s}\cdot \hat{p}'_s - \frac{1}{f\rho_s}\cdot\frac{\partial \hat{p}'_s}{\partial x\partial t}\ , \tag{8}$$

and

$$f\hat{v}'_s = \frac{1}{\rho_s}\frac{\partial \hat{p}'_s}{\partial x} - \frac{1}{f\rho_s}\cdot\frac{\partial^2 \hat{p}'_s}{\partial y\partial t}\ , \tag{9}$$

where ρ_s is the average density of water, $\beta = \frac{df}{dy}$ and f, the Coriolis parameter.

From the following relations

$$\rho_s = \rho_{s,0}(1-\alpha T_s),\quad \rho_s v = \rho_{s,0} v_0, \tag{10}$$

and

$$H^3\rho_{s,0}(1-\alpha T_s) = H_0^3\rho_{s,0}, \tag{11}$$

we have approximately

$$H = H_0\left(1+\frac{1}{3}\alpha T_s\right) \tag{12}$$

considering α is very small , where v is the volume of water with the length dimension of H, α, the thermal expansion coefficient of the sea water. Thus, the relations between the perturbing pressure p'_s, temperature T'_s and h, the perturbing height of mixed layer, are respectively

$$h = \frac{1}{3}D\alpha\hat{T}'_s\ , \tag{13}$$

and

$$\hat{p}'_s = \rho_s g h = \frac{1}{3}\rho_s g D\alpha\hat{T}'_s\ . \tag{14}$$

Eliminating u'_s and v'_s between Eqs. (8), (9) and (7) with the help of (14), we obtain the governing equation for variable h as follows

$$\left(\frac{\partial^2}{\partial x^2}+\frac{\partial^2}{\partial y^2}-\frac{f^2}{C_s^2}\right)\frac{\partial h}{\partial t} + \beta\frac{\partial h}{\partial x} + \frac{f}{\Delta\bar{T}_z}\left[\left(\frac{\partial \bar{T}}{\partial x}-\frac{\Delta\bar{T}_z}{D}\frac{\partial D}{\partial x}\right)\right.$$
$$\times\left(-\frac{\partial h}{\partial y}+\frac{\beta}{f}h+\frac{1}{f}\frac{\partial^2 h}{\partial x\partial t}\right)-\left(\frac{\partial \bar{T}}{\partial y}-\frac{\Delta\bar{T}_z}{D}\frac{\partial D}{\partial y}\right)$$
$$\left.\times\left(\frac{\partial h}{\partial x}-\frac{1}{f}\frac{\partial^2 h}{\partial y\partial t}\right)\right]=0\ , \tag{15}$$

where

$$C_s = \left(\frac{1}{3}gD\alpha\Delta\bar{T}_z\right)^{1/2} \tag{16}$$

is the speed of gravity wave in the mixed layer. In the equatorial region, Eq. (15) becomes approximately[8]

$$\left(\frac{\partial^2}{\partial x^2}+\frac{\partial^2}{\partial y^2}-\frac{\beta^2}{C_s^2}y^2\right)\frac{\partial h}{\partial t}+\beta\frac{\partial h}{\partial x}+\frac{f_0}{\Delta\overline{T}_z}\left[\left(\frac{\partial\overline{T}}{\partial x}-\frac{\Delta\overline{T}_z}{D}\frac{\partial D}{\partial x}\right)\right.$$

$$\times\left(\frac{\partial h}{\partial y}-\frac{\beta}{f_0}h+\frac{1}{f_0}\frac{\partial^2 h}{\partial x\partial t}\right)-\left(\frac{\partial\overline{T}}{\partial y}-\frac{\Delta\overline{T}_z}{D}\frac{\partial D}{\partial y}\right)$$

$$\left.\times\left(\frac{\partial h}{\partial x}-\frac{1}{f_0}\frac{\partial^2 h}{\partial y\partial t}\right)\right]=0 \quad , \tag{17}$$

where f_0 is the average value of f in the tropical region.

Introducing the non-dimensional variables

$$x=Lx^*,\quad y=\left(\frac{C_s}{\beta}\right)^{1/2}y^*,\quad t=(\beta C_s)^{-1/2}t^*, \tag{18}$$

where $(C_s/\beta)^{1/2}=\lambda_s$ is the radius of equatorial Rossby deformation, and $(\beta/C_s)^{-1/2}$ is the characteristic period of the Rossby wave. The non-dimensional form of Eq. (17) after dropping the asterisk is as follows

$$\left(\lambda^2\frac{\partial^2}{\partial x^2}+\frac{\partial^2}{\partial y^2}-y^2\right)\frac{\partial h}{\partial t}+\lambda\frac{\partial h}{\partial x}+S_x\left(\mu\frac{\partial h}{\partial y}+\lambda h\right.$$

$$\left.-\lambda^2\frac{\partial^2 h}{\partial x\partial t}\right)-S_y\left(\mu\frac{\partial h}{\partial x}-\frac{\partial^2 h}{\partial y\partial t}\right)=0 \;, \tag{19}$$

where

$$\lambda=\lambda_s/L,\quad \mu=f_0/\beta L, \tag{20}$$

and

$$S_x=\frac{\dfrac{\partial\overline{T}}{\partial x}-\dfrac{\Delta\overline{T}_z}{D}\dfrac{\partial D}{\partial x}}{\Delta\overline{T}_z}\;,\quad S_y=\frac{\dfrac{\partial\overline{T}}{\partial y}-\dfrac{\Delta\overline{T}_z}{D}\dfrac{\partial D}{\partial y}}{\Delta\overline{T}_z}\;. \tag{21}$$

For simplicity of mathematical treatment, the non-dimensional parameters S_x and S_y are taken to be constant.

III. DISPERSION RELATIONSHIP AND EIGENFUNCTION

Taking the solution of Eq. (19) as

$$h=\tilde{h}(y)\exp\left[-\frac{1}{2}\left(S_y+iS_x\frac{\mu}{\sigma}\right)y\right]\exp[i(kx-\sigma t)] \;, \tag{22}$$

Eq. (19) becomes

$$\frac{d^2\tilde{h}}{dy^2}+\left[-\frac{1}{4}\left(S_y+iS_x\frac{\mu}{\sigma}\right)^2-\left(\lambda^2k^2+\frac{\lambda k}{\sigma}-S_y\frac{\mu k}{\sigma}\right)\right.$$

$$\left.+iS_x\lambda\left(\frac{1}{\sigma}+\lambda k\right)-y^2\right]\tilde{h}=0 \;. \tag{23}$$

Obviously, if the relation

$$\frac{1}{4}\left(S_y+iS_x\frac{\mu}{\sigma}\right)^2+\left(\lambda^2k^2+\frac{\lambda k}{\sigma}-S_y\frac{\mu k}{\sigma}\right)-iS_x\lambda\left(\frac{1}{\sigma}+\lambda k\right)=-(2n+1) \tag{24}$$

is valid, the eigenfunction of Eq. (23) is

$$\tilde{h}(y)=\exp\left(-\frac{1}{2}y^2\right)H_n(y),\quad n=0,1,2,\cdots \tag{25}$$

where H_n is the Hermite polynomial with the order of n.

Rewritting (24), it becomes

$$\left[\lambda^2k^2+\frac{1}{4}S_y{}^2-(2n+1)-iS_x\lambda^2k\right]\sigma^2+\left[\lambda k-S_y\mu k-iS_x\lambda\right.$$
$$\left.+i\frac{1}{2}S_yS_x\mu\right]\sigma-\frac{1}{4}S_x^2\mu^2=0\ . \tag{26}$$

Two roots i. e. frequencies are obtained as follows

$$\sigma_{1,2}=\frac{1}{2}\left[\lambda^2k^2+\frac{1}{4}S_y^2+2n+1-iS_x\lambda^2k\right]^{-1}$$
$$\times\left\{-\left[\lambda k-S_y\mu k-iS_x\left(\lambda-\frac{1}{2}S_y\mu\right)\right]\pm\left[\left(\lambda k-S_y\mu k-iS_x\right.\right.\right.$$
$$\times\left.\left(\lambda-\frac{1}{2}S_y\mu\right)\right)^2+S_x^2\mu^2\left(\lambda^2k^2+\frac{1}{4}S_y^2\right.$$
$$\left.\left.\left.+(2n+1)-iS_x\lambda^2k\right)\right]^{1/2}\right\}\ . \tag{27}$$

In order to clarify the physical meaning of these waves obtained, two simplified cases are discussed as follows:

Case A. $S_x=S_y=0$, i. e. the influence of climatic background of ocean is not taken into account. In this case, there is only one root, i. e.

$$\sigma=-\frac{\lambda k}{\lambda^2k^2+2n-1}\ . \tag{28}$$

Obviously, it is the frequency of Matsuno's equatorial Rossby wave.

Case B. $S_x=0$, i. e. only the meridional temperature gradient and the slope of the top of thermocline are considered. In this case, the two roots of (27) degenerate into

$$\sigma_1=\begin{cases}0, & \mu S_y<\lambda,\\ \dfrac{(S_y\mu-\lambda)k}{\lambda^2k^2+\frac{1}{4}S_y^2+(2n+1)}, & \mu S_y\geqslant\lambda,\end{cases} \tag{29}$$

and

$$\sigma_2=\begin{cases}\dfrac{(S_y\mu-\lambda)k}{\lambda^2k^2+\frac{1}{4}S_y^2+(2n+1)}, & \mu S_y\leqslant\lambda,\\ 0, & \mu S_y>\lambda.\end{cases} \tag{30}$$

Comparing with (28), it can easily be understood that σ_2 is the modified Rossby wave, and σ_1 is another kind of wave, its propagating direction is opposite to the modified Rossby wave's. In general, μS_y takes positive values in the equatorial area so that the frequency of modified Rossby wave is reduced as S_y increases. Furthermore, when S_y is greater than its critical value, i. e.

$$S_{yc}=\frac{\lambda}{\mu}\ , \tag{31}$$

the modified Rossby wave disappears, and the new kind of wave becomes dominant. The dimensional form of (31) yields

$$\left(\frac{\partial \bar{T}}{\partial y}\right)_c-\frac{\Delta \bar{T}_z}{D}\left(\frac{\partial D}{\partial y}\right)_c=\frac{\Delta \bar{T}_z\beta}{f_0}=\frac{\Delta \bar{T}_z}{a\operatorname{tg}\theta}\ , \tag{32}$$

where θ is the latitude, a, the earth's radius. In the equatorial area of the Northern Hemisphere, in general, $\partial D/\partial y<0$, i. e. $\left(-\frac{\Delta \bar{T}_z}{D}\frac{\partial D}{\partial y}\right)>0$. Table 1 gives the critical meridional temperature gradient$\left(\frac{\partial \bar{T}}{\partial y}\right)_c$ for different latitudes and $\Delta \bar{T}_z$, in which $\frac{1}{D}\frac{\partial D}{\partial y}=-0.5\times 10^{-3}\ \mathrm{km}^{-1}$.

Table 1 Critical Values for Different θ and $\Delta \bar{T}_z$

$\Delta \bar{T}_z$(℃)		1	2	3	4	5
$\theta=5°$	$\left(\frac{\partial \bar{T}}{\partial y}\right)_c$(℃/$10^3km$)	1.29	2.58	3.87	5.16	6.45
$\theta=8°$	$\left(\frac{\partial \bar{T}}{\partial y}\right)_c$(℃/$10^3km$)	0.62	1.23	1.85	2.46	3.08

Returning the Eqs. (29) and (30) to the dimensional form and taking $\beta=0$, then the modified Rossby wave disappears and only the other kind of wave remains, i. e.

$$\sigma_1=\frac{\mu S_y k}{\lambda^2 k^2+\frac{1}{4}S_y^2+(2n+1)} \tag{33}$$

and

$$\sigma_2=0\,. \tag{34}$$

It shows that the wave with the frequency σ_1 is entirely due to the thermal structure of the sea temperature field. So, this kind of wave may be called the thermal wave. But it can only exist away from the equator by a short distance, at the equator where $f=0$, this kind of wave also disappears. In the middle latitudes, due to the fact that both $f\frac{\partial \bar{T}}{\partial y}$ and $-f\frac{\Delta \bar{T}_z}{D}\frac{\partial D}{\partial y}$ are negative, though this kind of wave does exist, its propagating direction is similar to the Rossby wave, i. e. westward. Essentially, the characteristic of this kind of wave, in the middle latitudes, is almost similar to the topographic Rossby wave indicated by Pedlosky[9]. Apparently, it corresponds to a new kind of wave in the tropical ocean.

IV. FREQUENCY ANALYSIS

On the basis of above discussions,we have known that two roots σ_1 and σ_2 corresponding to two kinds of waves exist in this model. One is the modified equatorial Rossby wave, and the other is the thermal wave mainly caused by the thermal structure of sea temperature in the mixed layer, i. e. S_x and S_y. The frequencies of these two kinds of waves in (27) are numerically calculated and the results are shown in Figs. 3—5 with $n=0$. The forms of curves for $n \neq 0$ are similar to those for $n=0$.

Fig. 3a shows the dispersion relationship between frequency and wavelength when S_y equals 0. The curve with $S_x=S_y=0$ is the equatorial Rossby wave obtained by Matsuno[10]. Fig. 3b shows the results with $S_y=0.25$, the solid and dashed lines corresponding to the modified Rossby and thermal waves respectively.

Fig. 4 shows the dispersion relationship when S_x takes the value -0.8. The calculated results are quantitatively about the same as the ones obtained by the simplified case B above. As $S_y<S_{y_c}=0.15$, there are modified equatorial Rossby waves with westward propagation while the eastward thermal waves appear in the domain of the parameters $S_y>S_{y_c}$. The most rapid propagating speeds of both waves are the ones with the wavelength of 2—3 thousand kilometers. Taking the parameters $\partial\bar{T}/\partial x=-3^\circ C/10^4$km, $\partial\bar{T}/\partial y=2.2^\circ C/10^3$ km, $1/D(\partial D/\partial x)=0.7/10^4$km and $1/D(\partial D/\partial y)=-2/3/10^3$km, the corresponding S_x and S_y are respectively -0.8 and 0.25. For these parameters, the periods and phase velocities of the modified Rossby wave and thermal wave are respectively $T_2=17$ years, $T_1=3.2$ months and $C_2=-10$ km/month, $C_1=646$ km/month for the wavelength 2×10^3 km and T_x 2.0°C.

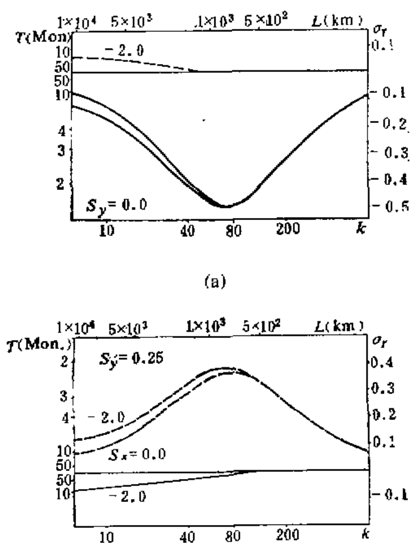


Fig. 3. Dispersion relationship for various values of S_x and S_y.
--- the thermal waves σ_{1y}
—— modified Rossby waves σ_{2y}

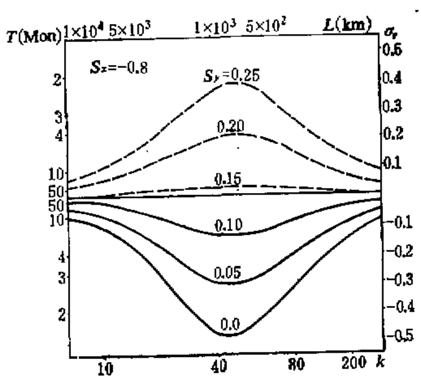
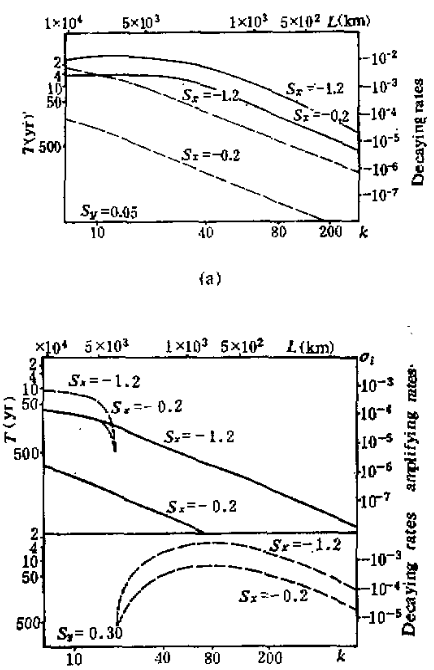


Fig. 4. Dispersion relationship of modified Rossby waves and thermal waves as $S_x=0.8$. The curve of σ_{1y} as $S_y<0.15$ and the curve of σ_{2y} as $S_y>0.15$ are not drawn owing to their very low frequencies. T oscillation period, L wavelength, k, σ_{1y} and σ_{2y} are non-dimensional wavenumber and frequencies respectively.

Figs. 5a and 5b show the amplifying and decaying rates of both waves for different values of S_x and S_y. For smaller S_y, say 0.05, both waves are decaying. For larger S_y, say 0.30, the modified Rossby wave is amplifying for all wavelengths while for the thermal waves, the situation appears that the longer waves are amplifying, the shorter waves are decaying, and the critical wavelength is approximately 3×10^3 km. Fig. 6 shows the amplifying and/or decaying rates with the values of S_y being 0.05 and 0.30 and $S_x=0.8$. Again, the longer thermal waves amplify and the shorter ones decay when S_y is 0.30, while the modified Rossby waves are amplifying for these parameters. Similarly, both waves are decaying for $S_y=0.05$. Whether the waves are amplifying or decaying, it takes time of the order of 10^0—10^2 years to reach the e-folding (or $1/e$) amplitude, so that these waves are almost neutral, indeed.



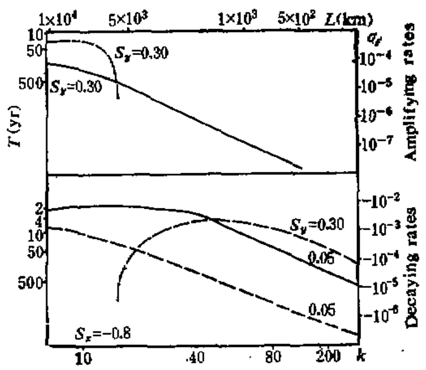


Fig. 5. The amplifying (or decaying) rates as $S_y=0.5$ and $S_y=0.30$. T is the e-folding time.

Fig. 6. Wave amplifying (or decaying) rates as $S_x=-0.8$. T is the e-folding time.

V. DISCUSSION

At the present stage, due to the scanty observed data it is impractical to explain the entire process of each ENSO, let alone developing a theory to account for the many observed aspects of the ENSO events such as those mentioned in section I. Nevertheless, it is possible to illustrate some main points in the ENSO events by means of the theoretical analysis, as is done in this paper. As well known, the easterlies over the Pacific would be anomalously strong in the year preceding the El Nino. The ocean response to the strong easterlies causes the SST to be much colder and the depth of mixed layer to be much shallower in the eastern part of the Pacific.This structure of sea temperature field will be maintained for a time even when the wind slackens due to the larger thermal inertia of the ocean. The situation of sea temperature field favorable for prevalence of eastward thermal waves has been obtained by this study. While one of the precusors of El Nino is an eastward displacement of the upward branch of the Walker Cell if the rising air is caused and/or accompanied by the warmer sea surface temperature, it is reasonable to suppose that, to a certain degree, the eastward displacement is caused by the effect of thermal waves. According to the calculation, the eastward propagating speed is about 640 km/month. If the upward branch of Walker Cell is associated with the thermal wave to make a displacement from Indonesia to the central area of the Pacific, it needs a time about one year. The thermal wave can continuously propagate eastwards until the westward trade-winds decrease very much, i. e., if the environmental thermal conditions are still favorable for prevailing of the thermal wave, then the thermal wave can arrive at the east part of the Pacific, and a modified equatorial Rossby wave may be excited there. Due to the slackening of the trade-winds the upwelling becomes weaker, the anomalously warm SST starts to develop in the eastern tropical Pacific Ocean, other conditions being suitable. If so, the modified Rossby wave carries the warmer anomalous SST westward, subsequently the warmer anomalous SST spread over east and central area of the Pacific, then the mature phase of El Nino is set up.

The main argument of our investigation is that the effect of the Kelvin wave in Wyrtki's theory is replaced by the thermal wave. As is known, the Kelvin wave can only exist on a line where the transverse velocities disappear no matter the medium is the atmosphere or the ocean. Whereas, the eastward propagating thermal wave only appear in a special environment, the most favorable environment being the tropical ocean due to the presence of a cold water pool. Thus we emphasize on the effect of thermal wave in the El Nino, a special phenomenon in tropical ocean.

However, there are some defects in explaining certain aspects of the ENSO events by using the thermal wave. A crucial problem is that the initial warmer anomalous SST should grow steadily to arrive at the exceptional high temperature in ENSO event,but the amplifying rates of thermal wave are too slow to satisfy the requirement of ENSO. Other effects should also be taken into account, for instance, the forcing action of the anomalous winds. The more important processes, are probably the anomalous air-sea interaction. Otherwise, we could not explain the southward displacement of the ITCZ in the eastern Pacific and the related phenomena in the atypical ENSO event.

Other processes associated with the thermal wave will be discussed in a separate report (Part II: linear numerical experiment, which will appear in the AAS Vol. 3, No, 2, 1986).

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