INSTABILITY OF THE OCEANIC WAVES IN THE TROPICAL REGION

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ABSTRACT

In this paper, the effects of the large-scale mean sea temperature fields of the tropical ocean and the zonal current field (southern equatorial current) have been comprehensively entered in consideration on the basis of Chao and Ji (1985), and Ji and Chao (1986), the equatorial oceanic waves of the tropical ocean have been discussed by use of linearized primitive equations, then, the significant influence of the climatic background fields of the tropical ocean upon the oceanic waves of this region has been further testified. When very cold water appears in the tropical region, and the southern equatorial current is also relatively strong, the effect of the Rossby wave weakens, as a consequence, there are substitutive slow waves (i.e. thermal waves) which travel in opposite direction (eastward) to the Rossby wave. The characteristics of the slow wave are similar to those of Rossby waves, only the travelling direction is opposite. Under a certain environmental background field, the slow wave and the modified Rossby wave may be instable. With this conclusion, the mechanism of the occurrence, development and propagation of El Nino events has been studied. It is pointed out that the opposite travelling direction of the thermal wave and Rossby wave will bring repectively into action under different marine environmental background fields. The physical causes for that the abnormal warm water inclines to occur along the South American coast have also been explored in this paper.

I. INTRODUCTION

At present, the studies on the air-sea interaction and El Nino/Southern Oscillation (ENSO) in the tropical areas have widely interested and drawn much attention in the community of meteorology and oceanography of the world. A lot of discussions about the propagation, extension of SST anomaly (SSTA) and the change of the thermocline have been undertaken. For example, Wyrtki (1975) pointed out that if trade winds weaken the Kelvin wave would be generated along the west coast of the Pacific, and when the Kelvin wave travels to the east coast, it would change the marine status of this region, i.e. the cold water upwelling weakens, then, SSTA increases correspondingly, as a consequence, the signal of the El Nino event appears. However, more scientists think (including Wyrtki. 1984) that ENSO is the results of the air-sea interaction.

Beyond doubt, it is very important to study the ENSO events from the point of view of air-sea coupling interaction. However, it is also necessary to study the dynamic characteristics of the marine itself and its effect in the El Nino events, because the marine dynamic process, in fact, has not been made completely clear. For instance, in the specific marine and mean field of the Pacific, the equatorial wave system is different from the Matsuno's results (1966). Using a filtered model of the shallow water wave of mixed layer with inertia-gravitational waves filtered out Chao and Ji (1985) and Ji and Chao (1986) have discussed the oscillation period and instability of the modified equatorial Rossby wave affected by a specific SST field in the Pacific through theoretical analyses and numerical experiments. They have pointed out that a group of thermal wave, which travels in opposite direction (eastward) to that of the Rossby wave may be generated in a certain conditon. The generation of this group of wave is caused by the horizontal inhomogeneity of SST, particularly by meridional gradients. The results of numerical experiment show that SSTA may propagate and increase with the aid of the eastward thermal waves and westward Rossby waves under the influence of the distribution of the specific equatorial mean SST field.

Based on the previous research (Chao and Ji, 1985; Ji and Chao, 1986), under the comprehensive consideration of the large-scale temperature fields and the equatorial current field, the equatorial wave and its instability have been further discussed theoretically with the shallow water model of the linearized primitive equation, and the physical factors of the occurrence and development of the El Nino events have been preliminarily studied with these results in this paper.

II. MODEL

The observational results show that the surface oceanic current of the Pacific in the equatorial region is the westward southern equatorial current. The position of this current has less change at the equator all the year round, only its strength changes with seasons. In this paper, we have not discussed the climatic causes for the formation of the equatorial current system, but, we only regard the equatorial current as a constant and basic current and as a background field to study the characteristics of the equatorial waves.

Fig. 2 in the paper by Chao and Ji (1985) shows that in the tropical Pacific regions, particularly the eastern Pacific, there is a very cold water area, SST of the tropical region increases with the increase of latitudes, and the temperature gradients exist in the east-west direction. The discussion in the paper (Chao and Ji, 1985) has shown that the temperature gradients in the east-west direction in the Pacific have little influence on the period of the slow wave (modified Rossby wave). Thus, only the influence of the gradients in the south-north direction of SST upon the tropical waves has been considered, and for the convenience of mathematical treatment, the mean gradient field of SST is assumed not to change with time.

Let us take x-axis of the coordinates system along the equatorial and positive eastward, y-axis is vertical to the x-axis and positive northward. The z-axis is positive upward. Analogous to the treatment in the paper (Chao and Ji, 1985), the sea surface is regarded as the upper surface of the model, and the top of the thermocline as the lower surface of the model.

The equation of continuity for the imcompressible fluid can be written as

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0, \tag{1}$$

then, we obtain approximately:

$$w(x,y,z,t)=-z\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)+\tilde{w}(x,y,t), \tag{2}$$

where $\tilde{w}(x, y, t)$ is an arbitrary function independent of z. The other symbols in this paper represent the same meaning as usual, except for special definition. On the sea surface at $z=h$, the normal velocity to this surface should equal zero because of the condition of vertical boundary. We thus obtain

$$w\Big|_{z=h}=u\frac{\partial h}{\partial x}+v\frac{\partial h}{\partial y}. \tag{3}$$

Inserting (3) into (2) obtains

$$\hat{w}_{S}(x,y,z,t)=(h-z)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)+u\frac{\partial h}{\partial x}+v\frac{\partial h}{\partial y}. \tag{4}$$

Thus, the perpendicularly-averaged vertical velocity $\hat{w}_S(x,y,t)$ is given by

$$\hat{w}_S(x,y,t)=\frac{D}{2}\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)+u\frac{\partial h}{\partial x}+v\frac{\partial h}{\partial y}, \tag{5}$$

where D is the chaacteristic depth from sea surface to the bottom of the mixed layer.

As an approximation, it may be assumed that the basic current is geostrophic. Although, strictly speaking this relation is not proper in some degree for the tropical regions, the diagnostic analysis of the dynamic height shows that in spite of the disturbed fluid field being a geostrophic, the mean ocean current in the tropical regions still have geostrophic features (Lukas and Firing, 1984). (Here, only the effect of the southern equatorial current is considered.) Therefore, we obtain

$$g\frac{\partial\bar{p}_S}{\partial y}=g'\frac{\partial\bar{h}}{\partial y}=-u_g f_0, \tag{6}$$

where $g'=\Delta\rho/\rho g$ is apparent gravitational acceleration, $\bar{h}$ is mean depth of mixed layer, $\partial\bar{p}_y/\partial y$ is mean meridional pressure gradient, and f_0 is mean value of Coriolis parameter near the equator. In fact, it is rational to take $f=\beta y$, however, there is no substantial influence on the results (Chao and Zhang, 1988).

Splitting disturbed and mean components, we have

$$u=u_g+u',\quad v=v',\quad \hat{w}_S=\hat{w}'_S,\quad h=\bar{h}+h'. \tag{7}$$

Inserting (6), (7) into (5) obtains

$$w_S'(x,y,t)=\frac{D}{2}\left(\frac{\partial u'}{\partial x}+\frac{\partial v'}{\partial y}\right)-\frac{u_g f_0}{g'}v'. \tag{8}$$

If there is no external heat sources, then the thermodynamic equation is

$$\frac{DT'}{Dt}+v'\frac{\partial\bar{T}}{\partial y}+\hat{w}'_S\frac{\Delta\bar{T}_z}{D}=0, \tag{9}$$

where T' is disturbed temperature, $\partial\bar{T}/\partial y$ is gradient in the north-south direction of the mean sea temperature field of the equatorial Pacific Ocean, $\Delta\bar{T}_z$ is temperature difference between sea surface and bottom of the mixed layer, and

$$\frac{D}{Dt}=\frac{\partial}{\partial t}+u_g\frac{\partial}{\partial x}. \tag{10}$$

In addition, the horizontal equations of motion are

$$\frac{Du'}{Dt}-\beta yv'=-G\frac{\partial T'}{\partial x}, \tag{11}$$

$$\frac{Dv'}{Dt}+\beta yu'=-G\frac{\partial T'}{\partial y}, \tag{12}$$

where $G=g\Delta\alpha$, and α is the coefficient of heat expansion.

Eqs. (8), (9), (11) and (12) constitute a completely closed equation system related to $u', v', \hat{w}'$ and T'.

III. SOLUTIONS OF THE EQUATIONS

Assuming that the form of the solutions are as follows

$$\begin{pmatrix} u' \\ v' \\ w' \\ T' \end{pmatrix} = \begin{pmatrix} u(y) \\ v(y) \\ w(y) \\ T(y) \end{pmatrix} e^{i(kx-\sigma t)}, \tag{13}$$

where k is wavenumber, σ is frequency. From Eqs. (8)—(13), we obtain

$$-i(\sigma-ku_g)u=-GikT+\beta yv, \tag{14}$$

$$-i(\sigma-ku_g)v=-G\frac{dT}{dy}-\beta yu, \tag{15}$$

$$-i(\sigma-ku_g)T+\frac{\Delta\overline{T}_z}{2}\left(iku+\frac{dv}{dy}\right)+S_y v=0, \tag{16}$$

where

$$S_y=\frac{\partial\overline{T}}{\partial y}-\frac{u_g f_0}{2Dg}. \tag{17}$$

The expression $\Delta\rho=\alpha\Delta\overline{T}_z\rho$ is used here. Putting

$$\tilde{S}_y=\frac{2S_y}{\Delta\overline{T}_z}, \qquad C_S^2=\frac{1}{2}gD\alpha\Delta\overline{T}_z. \tag{18}$$

From Eqs. (14)—(16) we obtain

$$\frac{d^2v}{dy^2}+b\frac{dv}{dy}+(c-dy^2)v=0, \tag{19}$$

where

$$\begin{cases} b=\tilde{S}_y, \\ c=\dfrac{(\sigma-ku_g)^2}{C_S^2}-\dfrac{ku_g(\sigma-ku)}{C_S^2}-\dfrac{[k(\sigma-ku_g)^2+\beta]k}{(\sigma-ku_g)}+\dfrac{k\tilde{S}_y f_0}{(\sigma-ku)}, \\ d=\dfrac{\beta^2}{C_S^2}. \end{cases} \tag{20}$$

By taking the following forms of transformation

$$v(y)=\tilde{v}(y)e^{-\frac{1}{2}\tilde{S}_y y}, \tag{21}$$

$$\left(\frac{\beta}{C_S}\right)^{1/2} y=y^*, \tag{22}$$

and omitting tildes and asterisks over variables, Eq. (19) then becomes

$$\frac{d^2v}{dy^2}+\left[\frac{c-\frac{1}{4}b^2}{d^{1/2}}-y^2\right]v=0. \tag{23}$$

To obtain Weber equation, the solution with boundary when $|y|\to\infty$ requires

$$\frac{c-\frac{b^2}{4}}{dy^2}=2n+1, \qquad n=0,1,2,\cdots. \tag{24}$$

The eigenfunction is as below

$$\tilde{v}(y^*)=e^{-\frac{1}{2}y^{*2}}H_n(y^*), \tag{25}$$

where H_n is Hermite's polynomial of n order.

The solution of the disturbed variable $v'(x,y,t)$ is

$$v'(x,y,t)=e^{-\frac{1}{2}\left(\frac{\beta}{C_S}y^2+\tilde{S}_y y\right)}e^{i(kx-\sigma t)}H_n\left[\left(\frac{\beta}{C_2}\right)^{1/2}y\right]. \tag{26}$$

At the same time, we can easily obtain the solutions of disturbed variables $fu'(x,y,t)$, $\hat{w}_S(x,y,t)$ and $T'(x,y,t)$. Here, we do not express one by one.

VI. INSTABILITY WAVES IN THE TROPICAL OCEAN

Now, let us discuss the period and the instability of the tropical wave. From Eq. (24) we can obtain

$$\omega^3-[k^2C_S^2+\frac{1}{4}\tilde{S}_y^2C_S^2+(2n+1)\beta C_S]\omega-(\beta-\tilde{S}_y f_0)C_S^2 k=0. \tag{27}$$

where

$$\omega=\sigma-ku_g. \tag{28}$$

1. *Brief Analysis of the Equation of Frequency*

If the effects of the mean current and large-scale sea temperature fields of the equatorial zone are ignored, then

$$\tilde{S}_y=0, \tag{29}$$

and Eq. (27) can be written as

$$\omega^3-[k^2C_S^2+(2n+1)\beta C_S]\omega-\beta C_S^2 k=0. \tag{30}$$

This is equation of dimensional form in the paper by Matsuno (1966). Similar to Matsuno's analysis, two fast waves (one traveling eastward and the other westward, i.e. inertia-gravitational waves) and the third one, equatorial Rossby wave traveling westward, have been obtained from Eq. (30). These three waves are all stable waves.

If $\tilde{S}_y\neq 0$, for the waves of relatively high wave numbers and high frequencies, we can approximately derive the following expression from Eq. (27)

$$\omega\doteq\pm\sqrt{k^2C_S^2+\tilde{S}_y C_S^2/4+(2n+1)\beta C_S}\,. \tag{31}$$

It may be obviously seen that these are still two inertia-gravitational waves with different traveling directions, only the frequencies are modified owing to the existence of $\tilde{S}_y$.

For the waves of low frequency, we approximately derive from Eq. (27) the following equation

$$\omega\doteq-\frac{(\beta-\tilde{S}_y f_0)C_S^2 k}{k^2C_S^2+\tilde{S}_y^2C_S^2/4+(2n+1)\beta C_S}\,. \tag{32}$$

The results are similar to those of Chao and Ji (1985). When the parameter $\tilde{S}_y$ is relatively small, thus $\omega<0$, the wave is the modified Rossby wave of the tropical region. When $\tilde{S}_y$ relatively large, $\omega>0$, this slow wave travels eastward, and we call it thermal wave.

From the discriminant of a cubic algebraic equation, we obtain the condition for conjugate complex root of Eq. (27) as

$$\Delta=\frac{(\beta-\tilde{S}_y f_0)^2}{4}K^2-\frac{[Kk+\tilde{S}_y^2C_S^2/4+(2n+1)\beta C_S]^3}{27}>0, \tag{33}$$

where

$$K=kC_S^2. \tag{34}$$

For the high wavenumber and relatively large value of $|\tilde{S}_y|$, we have obtained approximately the following expression

$$27|\tilde{S}_y|f_0/4>k^2C_S. \tag{35}$$

By use of wavelength $L=2\pi/k$, we obtain

$$L>\frac{C}{\sqrt{|\tilde{S}_y|}}, \tag{36}$$

where C is a constant greater than zero. Thus, when $|\tilde{S}_y|$ is relatively big and wavelength L satisfies Eq. (36), inevitably, there is one instable wave with growing amplitude among three roots of Eq. (27). It can be seen from Eq. (36) that when $\tilde{S}_y\ll 0$ and $\tilde{S}_y\gg 0$, two instable areas would appear correspondingly.

2. *Dispersion Relation and Growing Rate of Instable Wave*

When effects of the zonal basic circulation and the large-scale sea temperature field of the tropical regions are taken into consideration, usually, $\tilde{S}_y\neq 0$. From the numerical calculation of Eq. (27), we obtain dispersion relation and relation between growing rate and wavenumber.

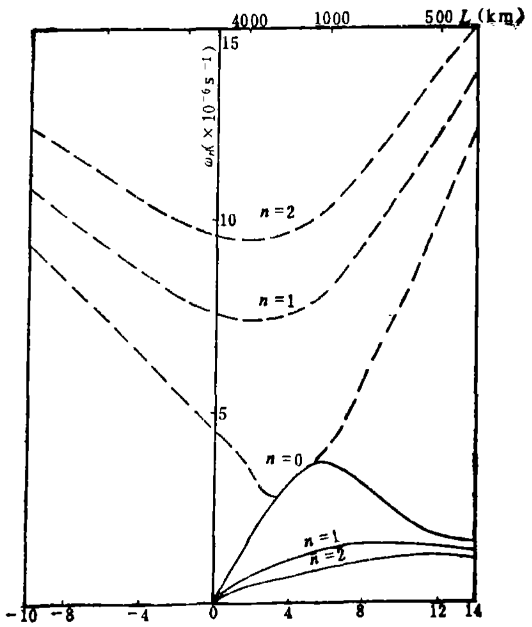

Fig. 1. Dispersion relation of the tropical oceanic wave when $\tilde{S}_y=3.3\times10^{-6}\text{m}^{-1}$. The dashed lines represent the inertia-gravitational wave, while the solid lines represent slow wave.

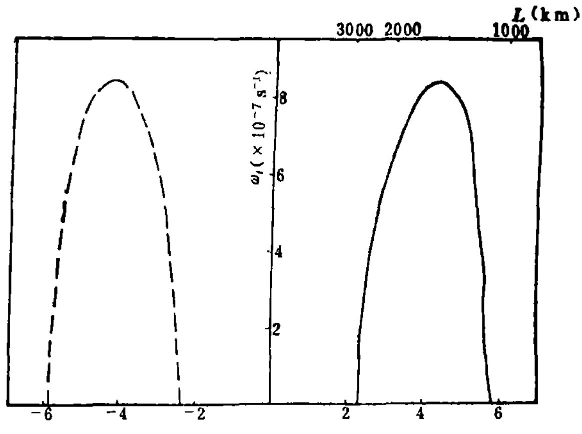

Fig. 2. Growing (decaying) rate, represented by solid (dashed) line, of the tropical oceanic wave when $\tilde{S}_y=3.3\times10^{-6}\text{m}^{-1}$.

Taking $\alpha=3\times10^{-4}\text{K}^{-1}$, $D=100\text{m}$, $g=9.8\ \text{m/s}^2$, $f_0=1.5\times10^{-5}/\text{s}$, $u_g=-0.1\ \text{m/s}$, $\partial T/\partial y=2.2\times10^{-6}\ {}^\circ\text{C/m}$, and $\Delta T_z=4.3^\circ\text{C}$, from Eq. (18), we can obtain $\tilde{S}_y=3.3\times10^{-6}/\text{m}$. At this time, there is a very cold water area in the equatorial zone. From Eq. (27), we obtain Fig. 1 of dispersion relation and Fig. 2 of relation between growing (decaying) rate and wavenumber. Comparing Fig. 1 with the results of Cane and Sarachik (1976) shows that when $n\geq 1$, though $\tilde{S}_y\neq 1$, the frequencies of two branches of inertia-gravitational waves have little changes. However, there is no Rossby wave traveling eastward (Cane and Sarachik, 1976), alternately, there is another slow wave traveling in the opposite direction to the Rossby wave. We call it thermal wave (Chao and Ji, 1985).

The generation of this slow wave is just due to the consideration of large-scale mean temperature field and mean current field of the tropical regions. When $n=0$, the mixed thermal-inertia-gravitational wave may be seen from Fig. 1. Fig. 2 shows that within the range of coincided frequencies of the inertia-gravitational wave with thermal wave, the waves are instable.

In Fig. 2, solid (dashed) lines represent the growing (decaying) rate of thermal-inertia-gravitational waves. The calculations show that when $n \geqslant 1$, the instability of the wave may appear if $|\tilde{S}_y|$ is certainly much greater. While the actual large-scale mean current field in the ocean is difficult to satisfy this condition. Therefore, only the case is given when $n=0$, in Fig. 2. Two branches of conjugate thermal-inertia-gravitaitonal mixed waves can be seen in Fig. 2, one is decaying wave, and the other intensified. The length of unstable waves lies between 2200—1280 km.

When the southern equatorial current of the equatorial zone weakens, even the eastward oceanic current appears (i.e. $u_g > 0$), and sea temperature has relatively large positive anomalies in this area, then it is known from Eq. (8) that $\tilde{S}_y < 0$. Fig. 3 shows dispersion relation of the waves in the tropical region when $\tilde{S}_y = -0.07 \times 10^{-6}$/m. Comparing Fig. 3 with the results of Cane and Sarachik (1976), it may be seen that when $n \geqslant 1$, frequencies of the Rossby wave and two branches of inertia-gravitational waves have little change. But, when $n=0$, it is known from Eq. (27) that the frequency relation of $\omega = -k$ does not exist if $\tilde{S}_y \neq 0$. Therefore, we have not taken away any root in Exp. (2), still, we can obtain the frequencies of three branches of waves. It may be seen from Fig. 3 that within the wavelength of 1.3×10^3km -2.2×10^3km, there are two waves i.e. Rossby-inertia-gravitational

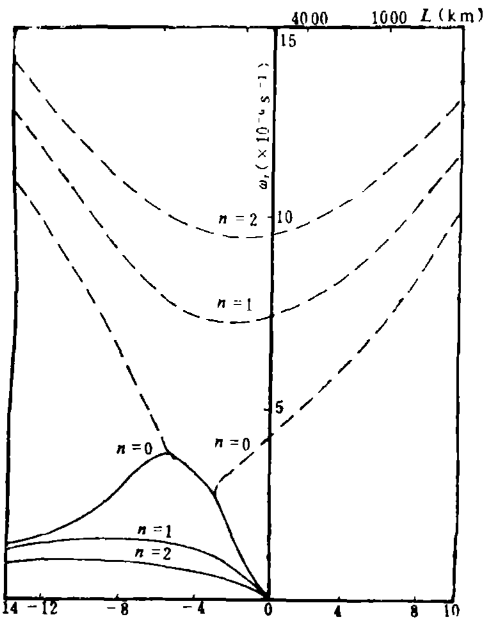

Fig. 3. Dispersion relation of the tropical oceanic wave when $\tilde{S}_y = -0.07 \times 10^{-6} m^{-1}$.

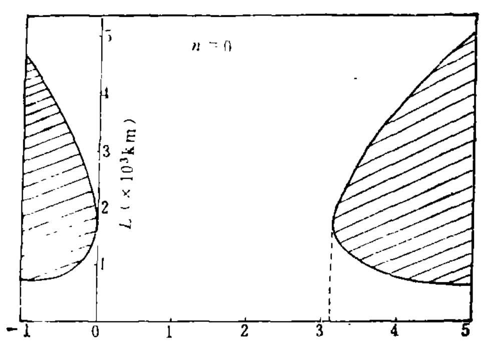

Fig. 4. Relation between parameter $\tilde{S}_y$ and instable length of oceanic wave. The area with the oblique lines is instable region.

mixed waves, with conjugate frequencies. The relation between their growing (decaying) rate and wave number k is similar to Fig. 2, we will not explain in this paper.

The calculations show that the frequency of the modified Rossby wave is gradually decreasing with the increase of $\tilde{S}_y$, when $\tilde{S}_y \geqslant S_{yc_1}=1.52\times10^{-6}$/m, its frequency trends to be zero. The alternative is thermal wave travelling eastward. Thus, when $\tilde{S}_y < S_{yc_1}$, the relation of dispersion obtained from Eq. (27) is similar to Fig. 3, while if $\tilde{S}_y > S_{yc_1}$, this relation is similar to Fig. 1, only with frequencies different (particularly for slow waves).

3. *Relation between Parameter $\tilde{S}_y$ and Instable Wavelength*

Fig. 4 shows the relation between $\tilde{S}_y$ and instable wavelength for $n=0$, which can be obtained by setting $\Delta=0$ in (33). Within the range of short wavelengths the two instable areas (hatched in Fig. 4) are generally similar to (36) approximately obtained above. From Fig. 4 we can see that when $\tilde{S}_y > 3.12\times10^{-6}$/m or $\tilde{S}_y < 0$, it is easy to generate instable waves near wavelength $L=1500$ km. Because the oceanic currents of the upper layer of the equatorial oceans are generally the southern equatorial currents travelling westward ($u_g<0$), when there is a cold water in the equatorial Pacific, the waves incline to enter into the right instable area in Fig. 4 according to (18). When a warm water area appears in the equatorial Pacific and the southern equatorial current weakens, the oceanic background field may satisfy the condition of $\tilde{S}_y < 0$, then the waves enter into the left instable area in Fig. 4.

V. INSTABLE WAVES AND EL NINO EVENTS

The occurrence and development of El Nino events and the propagation of SSTA always attract much interests and attention. Because ENSO is the result of very complicated air-sea coupling interaction, the detailedly theoretical studies on ENSO are not enough now. For the SSTA travelling westward, it is well known that this is owing to the effect of the Rossby wave, while for the SSTA travelling eastward, many researchers think this is the effect of the Kelvin wave that makes SSTA of the western Pacific travel eastward according to Wyrtki's theory (1975).

However, as Chao and Ji (1985) and Ji and Chao (1986) pointed out, the equatorial Kelvin waves certainly have influence on the El Nino events, but the dynamic condition of the existence of the equatorial Kelvin wave demands the current components in the south-north direction to be zero, in fact this condition could not be strictly satisfied in the current field of the equatorial Pacific, and certainly weakens the effect of the Kelvin wave. Ji (1987) also discussed this problem in his research on the response of the tropical atmosphere to the sea surface heat source. It is known from the above discussion that when $\tilde{S}_y > S_{yc_1}$, we may obtain eastward travelling thermal wave. In the El Nino events, the thermal wave may play the similar role to the Kalvin wave, and it makes SSTA propagate from the west to the east. In Fig. 1, if $L=2100$ km is given, the phase speed of the thermal wave is 0.95 m/s, i.e. it travels 10000 km in 130 days. This is approximated to the observational propagating speed of SSTA in the El Nino events. It should be pointed out that this speed is not sensitive to the change of parameters.

Although the air-sea interaction is not directly taken into account in this paper, only the physical characteristics of the ocean itself, i.e. the influence of the particular distribution of the large-scale mean SST field and current field in the equatorial Pacific on the oceanic waves (of course, these fields are the consequences of the long-range air-sea interaction) have been considered. Because instability exists in oceanic waves, it seems that this kind of instability may give a qualitative physical explanation to the mechanism of occurrence, development and propagation of the El Nino events. Fig. 4. shows that the equatorial waves are stable within the range of parameter $0<\tilde{S}_y<3.1\times10^{-6}$/m. Normally, actual large-scale mean current field (souhtern equatorial current) and SST field of the equatorial oceans confine the change of $\tilde{S}_y$ in this range. It seems that this may explain why large anomaly of SST does not exist. However if the equatorial trade wind persistently intensifies, it causes the southern equatorial current and the equatorial water upwelling to strengthen, and SST to decrease, thus leading to increase of parameter $\tilde{S}_y$. When $\tilde{S}_y$ exceeds the critical value, i.e. $S_{yc_2}=3.1\times10^{-6}$/m, the equatorial oceanic waves become instable, and the disturbed SST field increases instably. It may be seen from the curves of the growing rate in Fig. 2 that if $L=2100$ km, the e-fold growing time is 25 days. At the early stage of the El Nino events, SSTA may grow at this rate, at this moment, the thermal waves would play a dominant role and cause SSTA to propagate and to extend from west to east. While at the mature stage of the El Nino events, large areas of the warm water appear in the equatorial zone, the cold water areas even disappear. It is known from the analysis of Section IV, the modified Rossby waves play a dominant role at that time. Because $\tilde{S}_y<0$, the instable Rossby waves would travel westward. In this way, SSTA may complete an entire process, i.e. first, it propagates from west to east, then from east to west during El Nino events. The evolution and development of SSTA in the El Nino event of 1982—1983 are similar to this process.

During El Nino events, generally the positive SSTA appears first along the Southern American coast, i.e. the instably increasing of SST occurs first. The observations show that the cold water of the Southern American coast is the coldest. It is easy to satisfy the instable condition, $\tilde{S}_y>S_{yc_2}=3.1\times10^{-6}\text{m}^{-1}$ in Fig. 4 of this paper. It seems to explain qualitatively the physical reasons why abnormal SST inclines to occur along the Southern American coast at first. The conclusion drawn by Ji (1987) shows that the response of the atmosphere to the positive SSTA in Eastern Pacific may generate anomalous westerlies in the Middle and Western Pacific, furthermore, the anomalous westerlies are favourable to the further increase of SSTA.

Ji and Chao (1987) pointed out from their studies on the teleconnection of SST that during the early stage for each of El Nino events of 1975, 1963, 1969, 1972 and 1976, relatively strong negative SSTA may appear in the Eastern Pacific. We think that persistent decreasing of SST of the equatorial Eastern Pacific (i.e. the relatively large negative anomaly of SST appears) would tend to the increment of the gradients of the large scale sea temperature and makes $\tilde{S}_y$ increasing, thus the ocean waves enter into the left instable area in Fig. 4.

VI. DISCUSSION AND CONCLUSION

On the basis of the studies of Chao and Ji (1985) and Ji and Chao (1986), the effect of the large-scale mean temperature field of the equatorial ocean and zonal mean oceanic current have been taken into consideration, and the equatorial oceanic waves have been studied by use of linearized primitive equations in this paper. And the important influence of the climate background of the equatorial ocean on the tropical waves has been further verified. Under the effect of those climatic mean fields, the tropical waves may become instable. When $\tilde{S}_y > S_{yc_1} = 1.52\times10^{-6}\ \mathrm{m}^{-1}$, the Rossby waves disappear, the subsitutive eastward slow waves appear. We call them thermal waves (Chao and Ji, 1985). The characteristics of the thermal wave are similar to those of Rossby waves, only the traveling direction is opposite. When the cold water of the equatorial zone becomes colder and the southern equatorial current strengthens, it leads to $\tilde{S}_y > \tilde{S}_{yC} = 3.1\times10^{-6}\mathrm{m}^{-1}$, at this time the eastward thermal waves play a dominant role, and the instable phenomenon appears around the wavelength $L = 1500$ km. If the warmer water appears in the equatorial zone and the southern equatorial current weakens, then, $\tilde{S}_y < 0$, at that time the westward Rossby waves play a dominant role, and the instable phenomenon occurs around the wavelength $L = 1500$ km.

According to the conclusion above, the mechanism of the occurrence, development and propagation of the El Nino events have been discussed in this paper. It is pointed out that under the background fields of the different oceanic environments, the thermal wave and Rossby wave which travel in the opposite direction may play different roles respectively. The effect of the thermal wave is similar to that of Kelvin wave, i.e. it causes the instable SSTA to propagate from west to east. While the modified Rossby wave causes instable SSTA to propagate from east to west. Under the influence of these two slow waves, the whole propagating process of SSTA in an El Nino event may complete.

The physical causes for the abnormal warm water which tends to appear along the Southern American coast have been also discussed in this paper. There is a very strong equatorial cold water zone along the Southern American coast. When the trade wind intensifies, it cuases the ocean water upwelling and the Southern equatorial current to strengthen and the gradients of the large-scale ocean temperature to increase, furthermore, $\tilde{S}_y$ to increase. At this moment, the oceanic environmental background fields enter into instable area of $\tilde{S}_y > \tilde{S}_{yC}$ of Fig. 3. Fig. 1 in the paper by Ji and Chao (1987) also illustrated that generally SST may greatly decreases at the early stage of the El Nino event.

El Nino event is the result of the complicated nonlinear air-sea coupling interaction. It is impossible to explain its whole process by linearized shallow water model in this paper. Discussed in this paper are only the instable triggering mechanism which may be generated by the large-scale oceanic background field and the influence of the field on the propagating process of the SSTA during the El Nino event. As for the problem of the interannual change of the large-scale oceanic current and SST, it is not the concernment in this paper. The further study of the equatorial waves and El Nino events should use the coupling air-sea interaction model. Now, we have already set up a simple analytic model of this kind for tropical regions (Ji and Chao, 1988), and will further develop and perfect coupling air-sea interaction models.

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