The Westerly Anomalies over the Tropical Pacific and Their Dynamical Effect on the ENSO Cycles during 1980–1994[①]

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ABSTRACT

In this paper, the zonal wind anomalies in the lower troposphere over the tropical Pacific during 1980–1994 are analyzed by using the observed data. The results show that during the formation of the 1982 / 83, 1986 / 87 and 1991 / 92 ENSO events, there were the larger westerly anomalies in the lower troposphere over the equatorial Pacific. Moreover, it is explained by using the correlation analyses that the westerly anomalies over the equatorial Pacific could cause the warm episodes of the equatorial central and eastern Pacific.

A simple air–sea coupled model is used to discuss theoretically the dynamical effect of the observed westerly anomalies of wind stress near the sea surface of the equatorial Pacific on the ENSO cycle occurred in the period of 1981–1983. It is shown by using the theoretical calculations of the equatorial oceanic Kelvin wave and Rossby waves responding to the forcing of the observed anomalies of zonal wind stress near the sea surface of the equatorial Pacific that the westerly anomalies of wind stress near the sea surface of the equatorial Pacific make significant dynamical effect on the ENSO cycles occurred in the period of 1982–1983.

Key words: Westerly anomalies, ENSO cycle, Kelvin wave, Rossby wave

I. INTRODUCTION

ENSO events can be considered as the most important phenomenon in the tropical air–sea interaction, especially in the equatorial Pacific. When an ENSO event occurs in the equatorial Pacific, severe climate anomalies will be caused in the globe, and severe drought, flood and cooling summer will occur in some regions of the world.

Due to the above-mentioned reasons, recently, meteorologists and oceanologists pay great attention to the studies on the regularity and physical mechanism of ENSO phenomena. According to the recent results studied by many meteorologists and oceanologists, ENSO phenomenon is considered not only as an event, but also as a cycle.

Bjerknes(1966) first proposed a hypothesis on the formation mechanism of ENSO event. Bjerknes pointed out that ENSO cycle is a result of air–sea interaction. This hypothesis plays an important role either in the atmospheric sciences or in the oceanic sciences. Later on, many researchers introduced frequently this hypothesis in their studies.

Concerning the role of the eastern and western Pacific on ENSO event, it, all along, is a controversial problem. In the previous study, ENSO phenomenon was always considered as

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one occurred in the equatorial eastern Pacific along the Perusian coast. However, the warming of the ENSO event occurred in 1982 / 83 did not begin from the equatorial eastern Pacific, but from the equatorial central Pacific. Later on, it was also found that during many ENSO cycles, the warming of the tropical Pacific also began from the equatorial central Pacific. These made scientists cannot but doubt the formation mechanism of ENSO cycle proposed by Bjerknes.

Because it was found that the warming processes of some ENSO events are not in agreement with the theory proposed by Bjerknes, scientists were interested greatly in the formation mechanism of ENSO event, especially in the dynamical mechanism of ENSO cycle. Many investigations on this aspect have been made. For example, McCreay (1983), McCreay and Anderson (1984) and Anderson and McCreay (1985) systematically investigated the physical mechanism of ENSO cycle. They put forward the effect of the equatorial oceanic waves on ENSO cycle from theoretical analyses and observational facts. Although the dynamical mechanism of ENSO cycle proposed by McCreay may explain the dynamical effect of equatorial oceanic waves in the process of ENSO cycle, it cannot explain the effect of the tropical western Pacific on the formation of ENSO cycle. Later on, Schopf and Suarez (1988) put forward a possible dynamical mechanism of ENSO cycle from the unstable interaction between the atmosphere and the ocean in the tropical Pacific. Being different from the theory on the unstable air–sea interaction in ENSO cycle proposed by Schopf and Suarez (1988), Philander et al. (1984), Yamagata (1985) and Chao and Zhang (1988) investigated the air–sea interaction in the tropical Pacific and pointed out that the interaction between the atmosphere and the ocean in the tropics may form the unstable air–sea coupling waves, and ENSO event may be due to these coupling waves.

The above–mentioned studies show that the warm state of the western Pacific warm pool is a necessary condition for the occurrence of ENSO event. Only when the oceanic heat content of the western Pacific warm pool is in anomalous warm state, then, it is possible that ENSO event occurs. However, in the years of 1973, 1983 and 1989, the sea temperature of the western Pacific warm pool was in anomalous high state, and the oceanic heat content of the warm pool was anomalously large in those years, respectively, but ENSO event did not occur in the next year of those years. From this, it may be seen that the warm state of the western Pacific warm pool is only one of the necessary conditions for the occurrence of ENSO event. The atmospheric condition should be another necessary condition. Therefore, the anomalous tropical atmospheric circulation during the formation process of ENSO event should be analyzed. In this paper, the zonal wind anomalies in the lower troposphere over the tropical Pacific during 1980–1994 and their effect on the 1982 / 83, 1986 / 87 and 1991 / 92 ENSO cycles are discussed by using the observed data published in the Report on Climate System by J. M.A (Monthly Report on Climate System, 1991–1995), and the observed data analyzed by ECMWF. Moreover, a simple air–sea coupling model is used to discuss the dyamical effect of the observed westerly anomalies of wind stress near the sea surface of the equatorial Pacific on the ENSO cycle occurred in the period of 1981–1983.

II. THE WESTERLY ANOMALIES IN THE LOWER TROPOSPHERE OVER THE PACIFIC AND THEIR EFFECT ON ENSO EVENTS

1. *The Westerly Anomalies over the Tropical Pacific during 1980–1994*

Figure 1 indicates the interannual variation of the normalized anomalies of the zonal wind at 850 hPa over the equatorial central and eastern Pacific during 1980–1994. It may be seen from Fig.1 that the interannual variation of the normalized anomalies of the zonal wind appeared as an oscillation. The large westerly anomalies appeared in the ENSO years, such as 1982 / 83, 1986 / 87, 1991 / 92, 1993 and 1994, while the large easterly anomalies appeared in the anti-ENSO years, such as 1984 / 85, 1988, 1989. This may explain that there is a good relationship between the occurrence of ENSO events and the westerly anomalies in the lower troposphere over the equatorial central and eastern Pacific.

In order to show clearly the westerly anomalies in the lower troposphere from the tropical western Pacific to the equatorial central and eastern Pacific during the formation process of ENSO events, the longitude-time cross-sections of the zonal wind anomalies at 850 hPa over the equatorial Pacific, the Northern and Southern Hemispheric tropical Pacific during the formation processes of the 1982 / 83, 1986 / 87 and 1991 / 92 ENSO events are drawn, respectively. Figs. 2(a)-2(c) are the longitude-time cross-sections of the zonal wind anomalies at 850 hPa during 1981-1983, averaged along 5°N-5°S, Eq.-10°S and Eq.-10°N, respectively. It may be seen from Figs.2(a)-2(c) that the westerly anomalies propagated eastward from the tropical western Pacific to the tropical central and eastern Pacific. In the late half year of 1981, the westerly anomalies appeared over the tropical western Pacific, and these westerly anomalies propagated eastward from the tropical western Pacific to the tropical central and eastern Pacific during the period from the beginning of 1982 to the winter of 1982, and the maximum westerly anomaly located at the area to the east of the date line in the winter of 1982. Moreover, these westerly anomalies were gradually intensified during the eastward propagation and reached to 6.5 m / s before the occurrence of the ENSO event, and the maximum westerly anomaly could reach 8.0 m / s in the mature period of the ENSO event.

Figs. 3(a)-3(c) are the longitude-time cross-sections of the zonal wind anomalies at 850 hPa for 1985-1987, averaged along 5°N-5°S, Eq.-10°S and Eq.-10°N, respectively. From these figures, it may be found that the westerly anomalies propagated eastward from the equatorial eastern Indian Ocean in the summer of 1986 and reached the area near the date line in May, 1987. Compared Fig.3 with Fig.2, it can be seen that the westerly anomalies over the tropical Pacific during 1986-1987 were much weaker than those during 1982-1983 and the maximum westerly anomaly was only 3.0 m / s. Therefore, the ENSO event occurred in

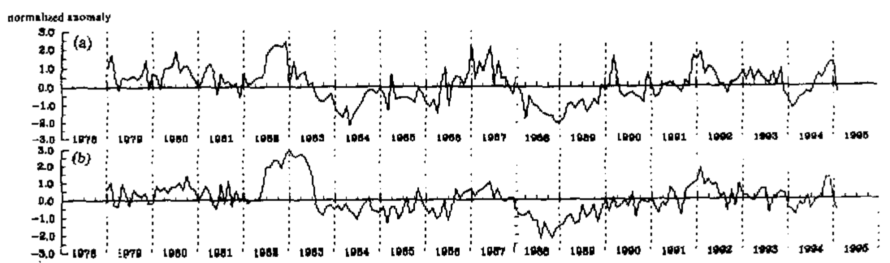


Fig.1. Interannual variation of the normalized anomalies of the zonal wind at 850 hPa over the equatorial central Pacific averaged for 165°E-175°W, 5°N-5°S (a) and over the equatorial eastern Pacific averaged for 175°E-135°W, 5°N-5°S (b), respectively.

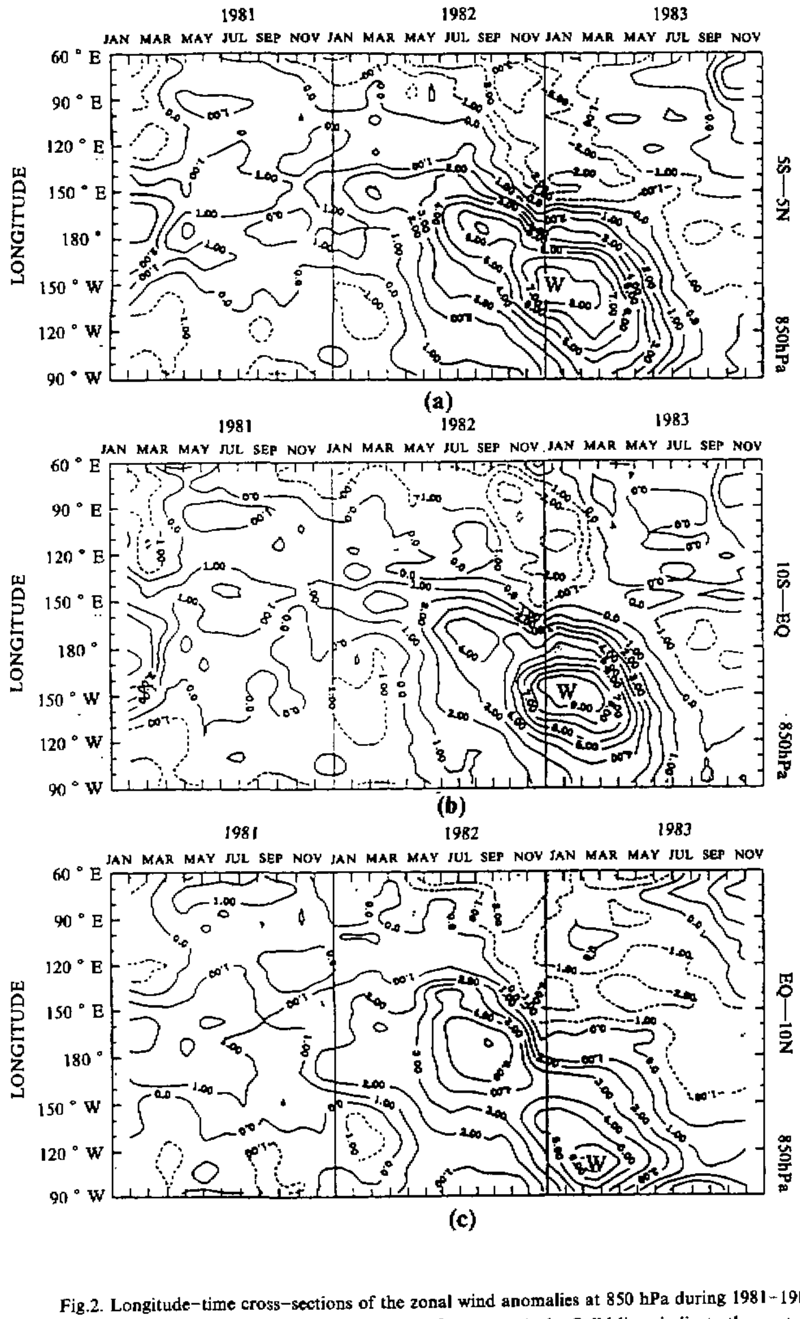


Fig.2. Longitude–time cross–sections of the zonal wind anomalies at 850 hPa during 1981–1983, averaged along 5°N–5°S, Eq.–10°S and Eq.–10°N, respectively. Solid lines indicate the westerly anomalies and dashed lines indicate the easterly anomalies, respectively. Units in m / s.

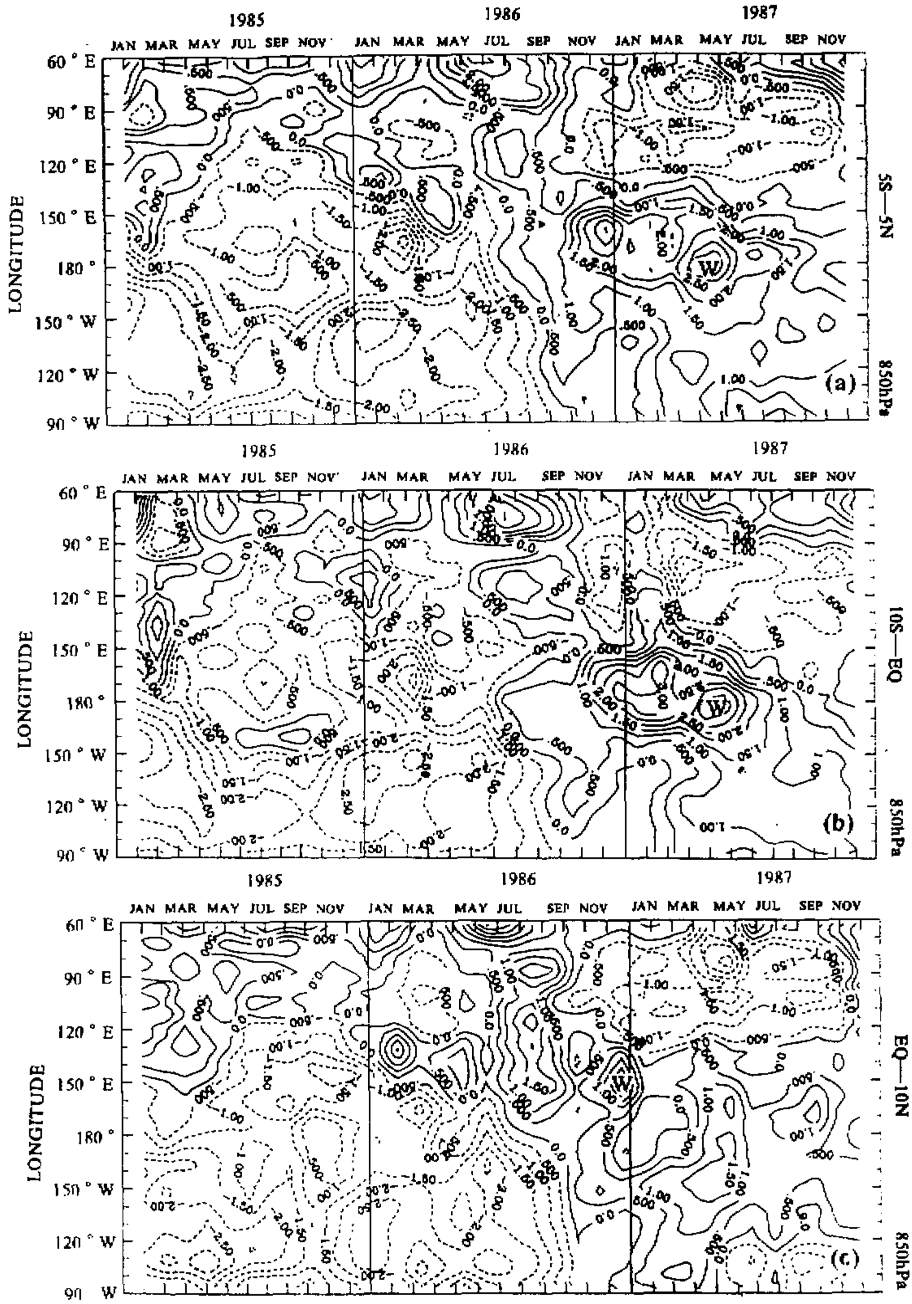


Fig.3. Same as in Fig.2, but for 1985–1987.

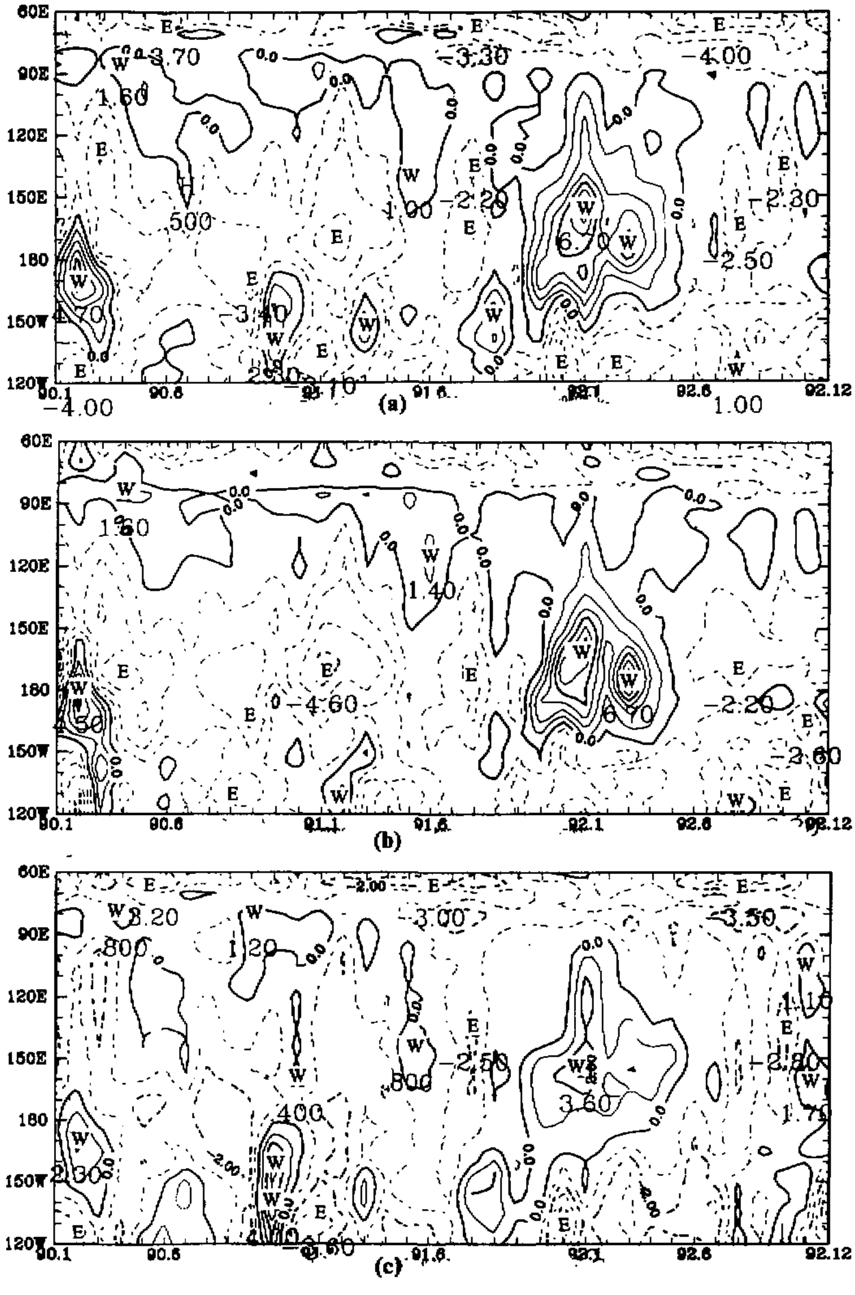


Fig.4. Same as in Fig.3, but for 1990–1992.

1986 / 87 was much weaker than that occurred in 1982 / 83. Moreover, the eastward propagation of the westerly anomalies during the formation process of the 1986 / 87 ENSO event was weak compared with the 1982 / 83 ENSO event. During the formation process of the 1986 / 87 ENSO event, the westerly anomalies propagated only to the equatorial central Pacific.

Figs. 4(a)–4(c) are the longitude–time cross–sections of the zonal wind anomalies at 850 hPa for 1990–1992, averaged along 5°N–5°S, Eq.–10°S and Eq.–10°N, respectively. It may be also found from Figs. 5(a)–5(c) that the westerly anomalies propagated eastward from the tropical western Pacific to tropical central and eastern Pacific during the formation process of the 1991 / 92 ENSO event, but the westerly anomalies suddenly increased before the onset of the ENSO event.

Obviously, it may be also seen from Figs. 2–4 that there were the easterly anomalies in the areas behind the westerly anomalies. Following the eastward propagation of the westerly anomalies, the easterly anomalies propagated eastward and extended to the equatorial central and eastern Pacific. The intensification of these easterly anomalies made the warm sea water be accomplished in the tropical western Pacific and the upwelling of the sea water be intensified in the tropical central and eastern Pacific. Thus, these easterly anomalies made the sea temperature decrease in the tropical central and eastern Pacific and made the ENSO event decay.

2. *Influence of the Westerly Anomalies over the Tropical Western Pacific on the SST Anomalies in the Equatorial Central and Eastern Pacific*

In order to investigate the influence of the westerly anomalies over the tropical western Pacific on the warming of the equatorial central and eastern Pacific, the simultaneous and lagged correlations between the westerly anomalies at 850 hPa, averaged for the area of 140°E–160°E, 5°N–5°S, and the SST anomalies in the Pacific Ocean are calculated by using the observed data for 36 months from 1981 to 1983.

Figs. 5(a)–5(c) show that the simultaneous and lagged correlations between the monthly mean westerly anomalies over the equatorial western Pacific and the monthly mean SST anomalies in the Pacific Ocean for 1981–1983. Figure 5 (a) indicates the simultaneous correlations between the westerly anomalies and the SST anomalies in the Pacific Ocean. From Fig.5 (a), it may be seen that positive correlation regions are located in the northern and southern subtropical Pacific and the equatorial central Pacific, while negative correlation is located in the equatorial eastern Pacific. Figs. 5 (b) and 5(c) indicate the lagged correlations between the westerly anomaly and the lagged SST anomalies in the Pacific Ocean for 6 and 10 months, respectively. It may be found that the correlation coefficients greatly change with the lagged time. The positive correlation coefficients located in the northern and southern subtropical Pacific gradually decrease with the lagged time and become negative up to the lagged time of six months. However, the negative correlation coefficients located in the equatorial eastern Pacific gradually become positive and the correlation coefficient reaches more than 0.5 in the lagged time of 10 months, and this positive area links with the area of positive correlation located in the equatorial central Pacific. This resembles the distribution of the SST anomalies during the mature period of ENSO events. Besides, the area of negative correlation located in the tropical western Pacific gradually becomes large and the correlation coefficients also increase.

From the distribution of simultaneous and lagged correlation coefficients it may be also seen that due to the continuous maintenance of the westerly anomalies over the tropical

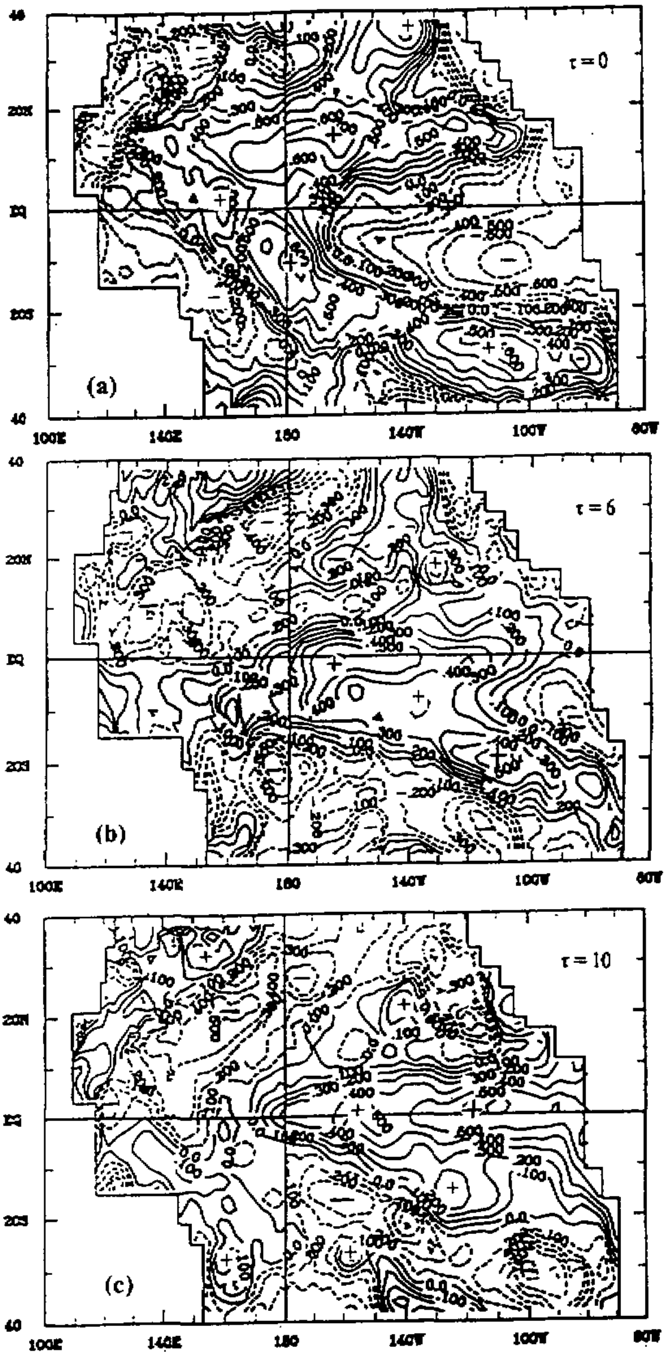


Fig.5. Simultaneous and lagged correlations between the monthly mean westerly anomaly at 850 hPa averaged for 140°–160°E, 5°N–5°S and the monthly mean SST anomalies in the Pacific Ocean. (a) Simultaneous correlation, (b) and (c) indicate the distributions of the correlation coefficients between the westerly anomaly and the lagged SST anomalies for 6 and 10 months, respectively. Solid lines indicate positive correlations and dashed lines indicate negative correlations.

western Pacific,the warm sea water propagates from the area near the dateline to the equatorial eastern Pacific.

Similarly, during the 1986 / 87 and 1991 / 92 ENSO cycles, the westerly anomalies over the tropical western Pacific also play an important role in the warming of the equatorial central and eastern Pacific.

III. A SIMPLE MODEL COUPLED BETWEEN THE ATMOSPHERE AND OCEAN IN THE TROPICAL PACIFIC

The dynamical effect of the westerly anomalies over the tropical Pacific on the formation of ENSO event has been a significant issue in the recent studies on the dynamics of ENSO cycle. Philander (1981) explained the effect of the relaxation of the anomalies of trade wind occurring in the equatorial central and eastern Pacific on the warming of the equatorial eastern Pacific with numerical experiments. Tang and Weisberg (1984) studied the influence of the anomalies of westerly wind stress during 1982–1983 on the warming of the equatorial central and eastern Pacific with a simple linear reduced–gravity model. However, this effect of the westerly wind stress on the warming of the equatorial central and eastern Pacific was discussed only by using numerical experiments or theoretical analyses, the dynamical effect of the observed westerly anomalies on the SST aomalies in the equatorial Pacific during an ENSO cycle was not yet discussed so far. Therefore, it is necessary to investigate further the dynamical effect of the actual westerly anomalies of wind stress occurring in the area near the sea surface of the equatorial Pacific on the oceanic waves in the equatorial Pacific and ENSO cycle.

In order to study the dynamical effect of the westerly anomalies of wind stress occurring in the area near the sea surface of the equatorial Pacific on the oceanic waves in the equatorial Pacific, the equatorial β–approximation, linear equations of shallow–water wave are applied in this study, and it may be written as follows:

$$\frac{\partial u}{\partial t} - \beta_0 y v = -g' \frac{\partial h}{\partial x} + \frac{\tau^x}{\rho H_0}, \tag{1}$$

$$-\beta_0 y u = -g' \frac{\partial h}{\partial x} + \frac{\tau^y}{\rho H_0}, \tag{2}$$

$$\frac{\partial h}{\partial t} + H_0 \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0, \tag{3}$$

where, u and v are the zonal and meridional perturbation velocities of the tropical Pacific current, respectively. h is the perturbation height of sea level, $g' = g\ \frac{\Delta\rho}{\rho}$ is the reduced gravity and it is taken as 5.6×10^{-2} m / s^2, $\Delta\rho$ is the density difference between the thermoclinic layer and the mixing layer of the tropical Pacific. H_0 is the sea level height without perturbation and may be taken as 150 m, $\rho = 1.026 \times 10^3$ kg / m^3, $\beta_0 = 2.28 \times 10^{-11}$s^{-1}m^{-1}, τ^x and τ^y are the anomalies of zonal and meridional wind stress near the sea surface of the tropical Pacific, respectively. $\vec{\tau} = (\tau^x, \tau^y)$ may be indicated as follows:

$$\vec{\tau} = \rho_a C_D (|\vec{\mathbf{V}}|\vec{\mathbf{V}} - |\overline{\mathbf{V}}| \cdot \overline{\mathbf{V}}), \tag{4}$$

where, $\rho_a = 1.275$ kg / m^3, $C_D = 3.2 \times 10^{-3}$, $\vec{\mathbf{V}}$ is actual wind field, and $\overline{\mathbf{V}}$ is climatological–mean wind field.

Nondimensionalizing Eqs. (1)–(3), and introducing new variables q and r (see Gill, 1980), i. e.,

$$q = h + u, \quad r = h - u, \tag{5}$$

where h and u are nondimensionalized variables, then, the obtained equations may be solved by using the Weber function,

$$\psi_m(y) = (2^m \ m! \ \pi^{\frac{1}{2}})^{-\frac{1}{2}} \exp\left(\frac{y^2}{2}\right) \frac{d^m}{dy^m} \left[\exp\left(-y^2\right)\right](-1)^m . \tag{6}$$

Thus, the following three kinds of waves can be obtained from the equations including the new nondimensionalized variables:

(1) For the response of the Kelvin wave, the equation is

$$\frac{\partial q_0}{\partial t} + \frac{\partial q_0}{\partial x} = \int_{-\infty}^{\infty} X\Psi_0 dy. \tag{7}$$

(2) For the response of the Rossby waves, the equation is

$$(2m+1)\frac{\partial q_{m+1}}{\partial t} - \frac{\partial q_{m+1}}{\partial x} = \int_{-\infty}^{\infty} X\left\{m\Psi_{m+1} - \left[m(m+1)\right]^{\frac{1}{2}}\Psi_{m-1}\right\}dy$$

$$+ 2\sqrt{m+1}\left(\frac{\partial}{\partial t} - \frac{\partial}{\partial x}\right)\int_{-\infty}^{\infty} Y\Psi_m dy, \qquad (m \neq 0) \tag{8}$$

and

$$r_{m-1} = q_{m+1}\left(\frac{m+1}{m}\right)^{\frac{1}{2}} - \left(\frac{2}{m}\right)^{\frac{1}{2}}\int_{-\infty}^{\infty} Y\Psi_m dy. \qquad (m \neq 0)$$

(3) For the response of the Yanai wave, i. e., the mixed Rossby–gravity wave, the equation is

$$q_1 = \sqrt{2}\int_{-\infty}^{\infty} Y\Psi_0 dy. \tag{9}$$

In the above–mentioned equations, X and Y are respectively the zonal and meridional nondimensionalized anomalies of wind stress near the sea surface of the equatorial Pacific. From the above–mentioned equations, we have the following conclusions:

1. Responding to the forcing of the anomalies of zonal wind stress near the sea surface, both the Kelvin wave and the Rossby waves may be excited in the equatorial ocean.

2. Responding to the forcing of the anomalies of meridional wind stress near the sea surface of the equatorial Pacific, the Rossby waves and the Yanai wave may be excited in the equatorial ocean, but the Yanai wave is limited only in the forcing area and cannot propagate.

For simplicity, only the dynamical effect of the anomalies of zonal wind stress near the sea surface of the equatorial Pacific on the equatorial oceanic waves is considered in the study. Obviously, the Yanai wave cannot be excited in this case. Moreover, it is assumed that the response of the equatorial oceanic waves to the anomalies of wind stress is symmetric to the equator, and the response of the Rossby waves is taken only up to the wave of four orders, then, the following equations can be obtained from Eqs. (7)-(9),

$$\frac{\partial q_0}{\partial t}+\frac{\partial q_0}{\partial x}=\int_{-\infty}^{\infty} \mathrm{X}\Psi_0 dy, \tag{10}$$

$$3\frac{\partial q_2}{\partial t}-\frac{\partial q_2}{\partial x}=\int_{-\infty}^{\infty} \mathrm{X}\left\{\Psi_2-\sqrt{2}\Psi_0\right\}dy, \tag{11}$$

$$7\frac{\partial q_4}{\partial t}-\frac{\partial q_4}{\partial x}=\int_{-\infty}^{\infty} \mathrm{X}\left\{3\Psi_4-2\sqrt{3}\Psi_2\right\}dy, \tag{12}$$

$$r_2=2\sqrt{\frac{1}{3}}q_4, \tag{13}$$

where X is as follows:

$$\mathrm{X}=\frac{\tau^x}{\rho H_0\ (C^3\beta_0)^{\frac{1}{2}}}. \tag{14}$$

Thus, the Kelvin wave and the Rossby waves responding to the observed anomalies of zonal wind stress near the sea surface can be solved from Eqs. (10)-(12).

IV. THE DYNAMICAL EFFECT OF THE WESTERLY ANOMALIES ON THE EQUATORIAL OCEANIC WAVES DURING THE 1982 / 83 ENSO CYCLE

In Section 2, the effect of westerly anomalies in the lower troposphere over the equatorial western and central Pacific on the warming of the equatorial central and eastern Pacific has been analyzed by using the observed data. In this section, the dynamical effect of the observed westerly anomalies of wind stress near the sea surface on the 1982 / 83 ENSO cycle will be explained by using the Kelvin wave and the Rossby wave responding to the observed anomalies of zonal wind stress, solved from Eqs. (10)-(12).

Figure 6 is the longitude-time cross-section of the nondimensionalized anomalies of zonal wind stress near the sea surface averaged for 5°N-5°S during 1981-1983, calculated by using formula (14). As shown in Fig.6, before the onset of the 1982 / 83 ENSO event, the easterly anomalies of wind stress were located in the area near the sea surface of the equatorical central and eastern Pacific, while there were the weaker westerly anomalies of wind stress in the area near the sea surface of the equatorial western Pacific. Moreover, the weaker westerly anomalies of wind stress located in the area near the sea surface of the equatorial western Pacific were intensified in the area near the sea surface of the equatorial central Pacific and moved eastward from March, 1982. In December, 1982, the westerly anomalies of wind stress reached maximum in the area near 120°W, and were maintained up to the beginning of 1983. After the beginning of 1983, the westerly anomalies of wind stress located in the area near the sea surface of the equatorial central and eastern Pacific gradually weakened and turned into the easterly anomalies. The anomalies of zonal wind stress were corresponding to the anomalies of zonal wind shown in Fig.1.

Thus, the Kelvin wave and the Rossby wave in the equatorial Pacific may be calculated by using Eqs. (10)-(12) and the anomalies of zonal wind stress shown in Fig.6. The lower figures in figure 7 show the observed SST aomalies in the equatorial Pacific and the observed nondimensionalized anomalies of zonal wind stress near the sea surface of the equatorial Pacific, respectively, and the upper figures indicate the Kelvin wave and the Rossby waves of

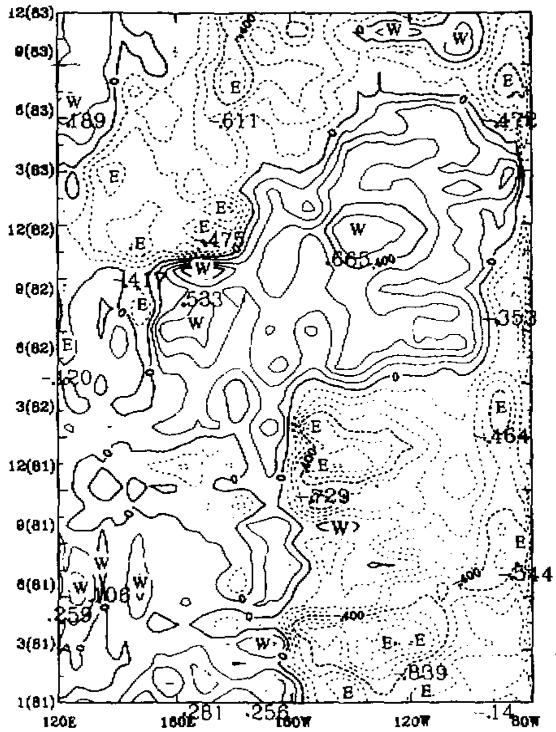


Fig.6. Longitude-time cross-section of the nondimensionalized anomalies of zonal wind stress at 1000 hPa averaged for 5°N-5°S during 1981-1983.

two and four orders in the equatorial Pacific responding to the observed anomalies of zonal wind stress during 1982–1983. It may be found from the lower part in Fig.7(a) that during the beginning of 1982, the easterly anomalies of wind stress were located in the area near the sea surface of the equatorial central and eastern Pacific, and SST anomalies in the equatorial Pacific were like the distribution of SST anomalies during the episode of La Nina event, i.e., the SST anomalies were positive in the equatorial western Pacific and negative in the equatorial central and eastern Pacific, respectively. As shown in the upper part in Fig.7(a) that the eastward-propagating cold Kelvin wave responding to the easterly anomalies of wind stress appeared in the equatorial eastern Pacific, and it was reflected by the eastern boundary of the equatorial Pacific and became the westward-propagating cold Rossby wave. These waves made the SST in the equatorial eastern Pacific maintain the cold state, as shown in the lower part in Fig. 7(a). Moreover, it may be seen that the westward-propagating warm Rossby waves responding to the easterly anomalies of wind stress appeared in the equatorial central Pacific, and they propagated westward and caused the warming of the western Pacific warm pool. This is in agreement with the observed warm state of the warm pool, as shown by the dashed line in the lower part in Fig. 7(a).

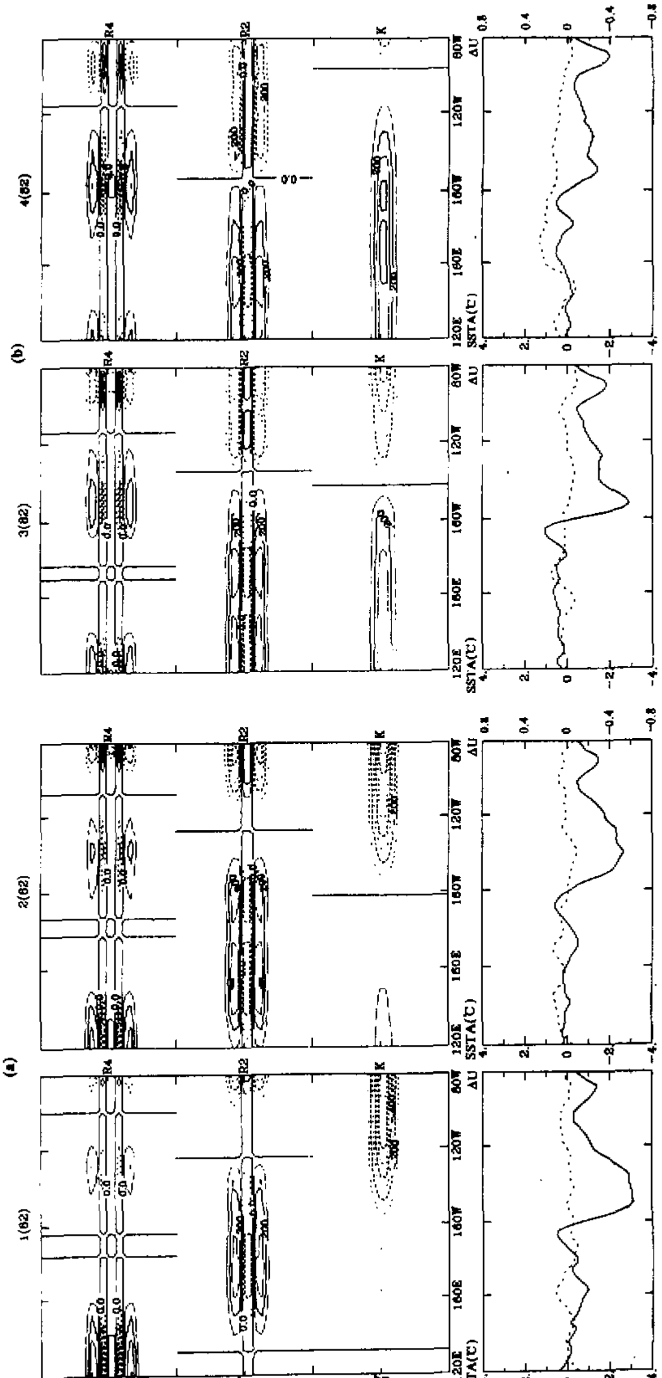
However, as shown in the lower part in Fig.7(b), during March-April of 1982, the westerly anomalies of wind stress appeared in the area near the sea surface of the equatorial western Pacific, due to the forced effect of this westerly anomalies, the warm Kelvin wave appeared in the equatorial western and central Pacific, and it propagated eastward. During May–June of 1982, the eastward–propagating warm Kelvin wave propagated into the equatorial central Pacific, and its amplitude increased there. This caused the warming of the equatorial central Pacific, as shown by the dashed line in the lower part in Fig.7(b). On the other hand, it may be found that due to the dynamical effect of the westerly anomalies, the cold Rossby waves can be excited in the equatorial central and western Pacific, and these waves propagated westward, and their amplitudes gradually increased during the westward propagation. These westward–propagating cold Rossby waves could cause the cooling of the western Pacific warm pool and provide the pre–condition of the formation of the cold Kelvin wave. With the eastward propagation of the westerly anomalies, as shown in the lower figure in Fig.7 (c), the strong westerly anomalies of wind stress appeared in the area near the sea surface of the equatorial central Pacific during September and Octorber of 1982. Thus, the eastward–propagating warm Kelvin wave responding to the forcing by the westerly anomalies moved into the equatorial central and eastern Pacific, and it was reflected by the eastern boundary of the equatorial Pacific and became the warm Rossby wave. Due to the interaction between the warm Kelvin and the reflected warm Rossby wave, the SST in the equatorial central and eastern Pacific greatly increased, as shown by the dashed line in the lower part in Fig.7 (c), the 1982 / 83 ENSO event occurred. Moreover, due to the maintenance of the westerly anomalies in the region near the sea surface of the equatorial central and eastern Pacific during the winter of 1982, the SST in the equatorial central and eastern Pacific continued to increase, as shown in the lower part in Fig.7(d), and the 1982 / 83 ENSO event reached the mature phase.

However, due to the eastward propagation of the cold Rossby wave, the western Pacific warm pool gradually became cold, as shown in the lower parts in Fig.7(c) and Fig.7(d), and the cold Kelvin wave was caused there and continuously propagated eastward. Moreover, the easterly anomalies appeared in the region near the sea surface of the equatorial western Pacific after the beginning of 1983, which made the cold Kelvin wave intensify greatly. With the eastward propagation of the easterly anomalies, this cold Kelvin wave gradually propagated eastward and propagated into the equatorial eastern Pacific after the spring of 1983. Thus, the SST in the equatorial eastern Pacific decreased, and the ENSO event decayed in the summer of 1983. By this way, the 1982 / 83 ENSO cycle including formation, developing, mature and decaying stages finished. Moreover, due to the unstable interaction between the atmosphere and the ocean in the tropical Pacific, the westward–propagating warm Rossby waves could appear in the equatorial central Pacific, and they could cause the warming of the western Pacific warm pool again. This provided a pre–condition for next ENSO cycle, i. e., the 1986 / 87 ENSO event.

In the upper parts of Fig.7 (a)–7(d), K, R_2 and R_4 indicate the Kelvin wave, the Rossby wave of two–order and four–order, respectively, and solid line shows a warm wave and dashed line shows a cold wave, the ordinate stands for the amplitude of these waves.

In the lower parts of Figs.7(a)–7(b), solid line indicates the observed nondimensionalized anomalies of zonal wind stress, and dashed line shows the SST anomalies in the equatorial Pacific.

Similarly, the Kelvin wave and the Rossby waves responding to the forcing by the observed westerly anomalies of wind stress near the sea surface of the equatorial Pacific during the 1986 / 87 and the 1991 / 92 ENSO cycles may be also calculated by using Eqs. (10)–(12) and the observed anomalies of zonal wind stress near the sea surface of the equatorial Pacific.


(a)
(b)
1(82)
2(82)
3(82)
4(82)
R4
R2
K
SSTA(℃)
AU
120E
160E
160W
120W
80W

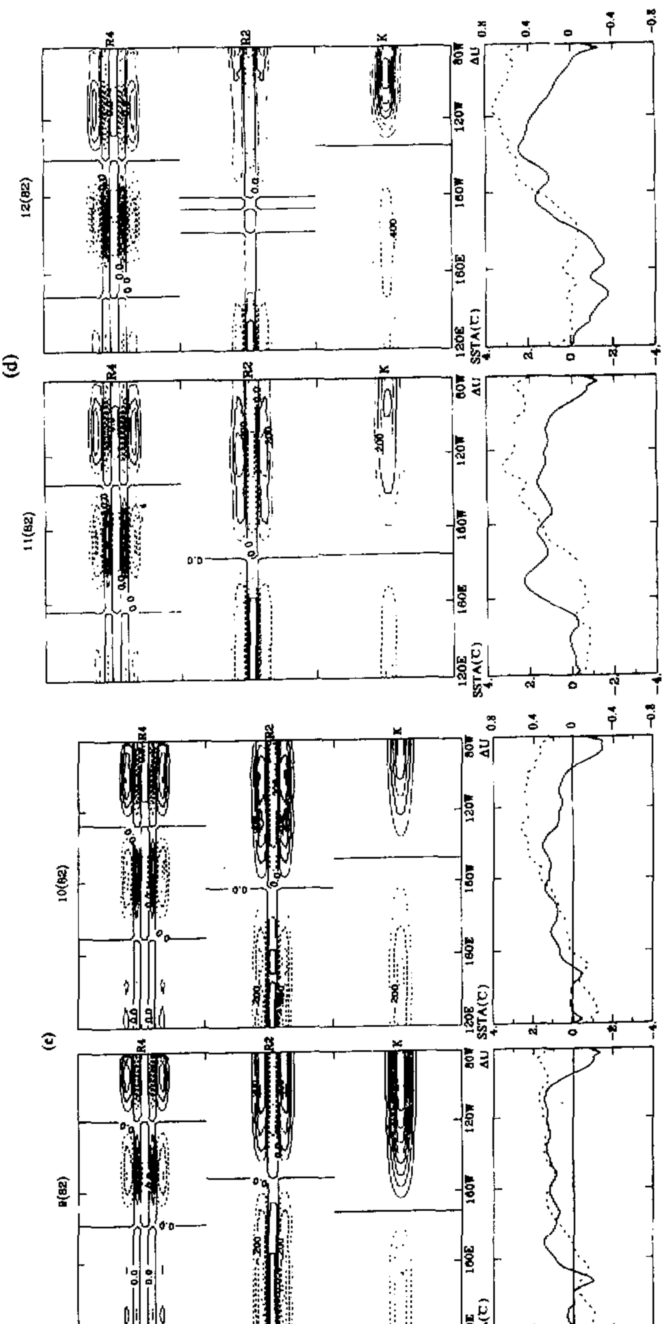


Fig.7. The temporal–spatial distributions of the anomalies of zonal wind stress near the sea surface, the SST anomalies and the Oceanic waves responding to the anomalies of zonal wind stress near the sea surface of the equatorial Pacific. (a) *in January and February*, 1982, (b) in March and April, 1982, (c) in September and October, 1982, (d) in November and December, 1982.

From the temporal and spatial evolutions of the Kelvin wave and the Rossby wave responding to the observed anomalies of zonal wind stress near the sea surface of the equatorial Pacific, it may be shown clearly that the westerly anomalies of wind stress near the sea surface of the equatorial Pacific make significant dynamical effect on ENSO cycle through the Kelvin wave and the Rossby wave excited by the westerly anomalies of wind stress near the sea surface of the equatorial Pacific.

V. CONCLUSIONS AND DISCUSSIONS

In this paper, the influence of the zonal wind anomalies in the lower troposphere over the equatorial Pacific on the ENSO cycles during 1980–1994 is analyzed by using the observed data. The analyzed results show that during these three ENSO events, there were the larger westerly anomalies in the lower troposphere over the equatorial Pacific, and the westerly anomalies over the equatorial Pacific could cause the warm episodes of the equatorial central and eastern Pacific. It may explain that the westerly anomalies over the equatoiral Pacific play significant role in the formation and development of the ENSO events. The westerly anomalies in the lower troposphere over the equatorial Pacific may be one of the necessary conditions for the formation of ENSO event.

A simple air–sea coupled model is used to discuss theoretically the dynamical effect of the observed westerly anomalies of wind stress near the sea surface of the equatorial Pacific on the ENSO cycles, especially on the 1982 / 83 ENSO cycle. From the theoretical solutions of the simple air–sea coupled model used in this paper, it is shown that under the effect of the westerly anomalies of wind stress near the sea surface of the equatorial western and central Pacific, the eastward–propagating warm Kelvin wave and the westward–propagating cold Rossby wave can be excited. The warm Kelvin wave responding to the forcing by the westerly anomalies of wind stress can propagate eastward and can be intensified with the eastward propagation of the westerly anomalies, and when the warm Kelvin wave propagates into the equatorial eastern Pacific, the ENSO event occurs. On the other hand, when the cold Rossby wave responding to the forcing by the westerly anomalies of wind stress propagates into the tropical western Pacific, the cold state of the western Pacific warm pool will be caused and the cold Kelvin wave can be exited. Under the effect of the easterly anomalies, the cold Kelvin waves will propagate eastward and can be intensified with the eastward propagation of the easterly anomalies, and it will made the ENSO event decay. By this way, an ENSO cycle will finish, and a new ENSO cycle will begin.

The above–mentioned results show that the westerly anomalies over the equatorial Pacific make significant effect on the ENSO cycles occurred in the period of 1980–1994. Where are the zonal wind anomalies from? Our study also illustrated that the westerly anomalies over the equatorial Pacific are associated not only with the unstable interaction between the atmosphere and the ocean in the tropics but also with the southward propagation of the westerly anomalies from the East Asian monsoon region, which may be connected with the EU pattern teleconnection of atmospheric circulation anomalies. The study will be published in another paper.

In this paper, only the dynamical effect of the zonal wind anomalies over the equatorial Pacific on ENSO cycle is discussed. However, the meridional wind anomalies may make certain dynamical effect on ENSO cycle. This needs to be investigated further.

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