Meridional Wind Stress Anomalies over the Tropical Pacific and the Onset of El Niño Part II: Dynamical Analysis①

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ABSTRACT

The data analyses in the first part of this study have shown that the sea surface temperature anomalies (SSTA) in the eastern equatorial Pacific are significantly correlated with the preceding anomalous convergence of the meridional wind stress near the equator. In order to understand the dynamical role of the convergent meridional wind stress anomalies in the El Niño occurring, an ideal wind stress which converges about the equator is set up based on the observations revealed in the first part. A simple dynamical model of tropical ocean is used to study the response of the tropical ocean to the convergent meridional wind stress. The results show that the convergent wind stress in the eastern equatorial Pacific is favorable for the occurrence of El Niño. When the convergent wind stress exerts on the tropical ocean, the westward propagating Rossby wave is excited, which, on the one hand, makes the mixed layer near the equator become thicker. On the other hand, the westward oceanic currents associated with the Rossby wave appear in the vicinity of the equator. The oceanic currents can drive the upper layer sea water to transfer to the west, which is favorable for the sea water to pile up in the western equatorial Pacific and to accumulate energy for the upcoming warm event.

Key words: Meridional wind stress, El Niño

1. Introduction

In the first part of this study (Zhang et al., 2001), by using the correlation and singular value decomposition (SVD) methods, the relations between the SSTA in the equatorial Pacific and the preceding meridional wind stress anomalies near the equator were diagnosed. It is found that prior to El Niño events by more than half a year, the meridional wind stress anomalies converge near the equator in the eastern equatorial Pacific. And it is inferred that the convergence of the meridional wind stress anomalies may be important in the occurrence of El Niño events. In order to investigate if there exists intrinsic relation between convergent meridional wind stress anomalies and the occurrence of El Niño events, in this paper we study the dynamical role of the convergent meridional wind stress anomalies over the tropical ocean by using a simple tropical oceanic model. In Section 2 the dynamical oceanic model is given. Sections 3 and 4 discuss the stationary and non–stationary responses of the tropical ocean to

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the convergent meridional wind stress anomalies, respectively. Summary and discussion follow in Section 5.

2. A simple dynamical model of the tropical ocean

A shallow water model on the equatorial β-plane is utilized in the present study. If only the meridional wind stress is considered, the model can be given as below:

$$\frac{\partial u}{\partial t}-\beta yv+g'\frac{\partial h}{\partial x}=-Au\ , \tag{1a}$$

$$\frac{\partial v}{\partial t}+\beta yu+g'\frac{\partial h}{\partial y}=-Av+\frac{\tau^y}{\rho H_0}\ , \tag{1b}$$

$$\frac{\partial h}{\partial t}+H_0\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)=-Bh\ , \tag{1c}$$

where g' is reduced gravity, H_0 is the undisturbed thickness of the oceanic mixed layer, ρ is the sea water density, τ^y is the meridional wind stress, A and B are the Rayleigh friction and Newtonian cooling coefficients, respectively, and the others are the same as the general definition.

By introducing the characteristic time scale $(2\beta c)^{-1/2}$ and space scale $(c/2\beta)^{1/2}$, where $c=(g'H_0)^{1/2}$, the non-dimensional form of Eqs.(1) can be written as

$$\frac{\partial u}{\partial t}-\frac{1}{2}yv+\frac{\partial h}{\partial x}=-\varepsilon u\ . \tag{2a}$$

$$\frac{\partial v}{\partial t}+\frac{1}{2}yu+\frac{\partial h}{\partial y}=-\varepsilon v+Y\ , \tag{2b}$$

$$\frac{\partial h}{\partial t}+\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}=-\gamma h\ , \tag{2c}$$

where $\varepsilon=A(2\beta c)^{1/2}$ and $\gamma=B(2\beta c)^{-1/2}$, the non-dimensional Rayleigh friction and Newtonian cooling coefficients, respectively, $Y=\tau^y/[\rho H_0(2\beta c^3)^{1/2}]$, the non-dimensional meridional wind stress. Eqs.(2) are the basic equations for the following analysis.

3. Stationary response of the equatorial ocean to convergent meridional wind stress

At first, in this section we will consider the simplified case, i.e., assuming that the equatorial ocean is stationary and the convergent meridional wind stress is zonally distributed evenly.

3.1 *Simplification and resolution*

As pointed by Charney and Flierl (1981), the dissipation terms $-\varepsilon u$, $-\varepsilon v$ and $-\gamma h$ are important in the formation of stationary oceanic currents. Yamagata and Philander (1985) discussed relative importance of the Rayleigh friction and Newtonian cooling. It is pointed out that the Newtonian cooling $-\gamma h$ is of central importance in the formation of stationary oceanic currents in the tropical ocean. Therefore, we reserve the dissipation terms stationary oceanic currents in the tropical ocean. Therefore, we reserve the dissipation terms and, for simplicity, assume $\varepsilon = \gamma$ in Eqs.(2). By dropping the time–dependent terms in Eqs.(2), we get the equations for stationary oceanic response to the meridional wind stress as below:

$$\varepsilon u - \frac{1}{2} y v + \frac{\partial h}{\partial x} = 0 \ , \tag{3a}$$

$$\varepsilon v + \frac{1}{2} y u + \frac{\partial h}{\partial y} = Y \ , \tag{3b}$$

$$\varepsilon h + \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0 \ . \tag{3c}$$

In the resolution of Eqs.(3), we take the natural boundary conditions in both x and y directions. According to the analyses in the first part of this study (Zhang et al., 2001), the meridional wind stress Y can be simply considered as

$$Y = - D_1(y) \ , \tag{3'}$$

where $D_1(y)$ is the 1st–order parabolic cylinder function and $D_1(y) = y\exp(-y^2/4)$ (Wang and Guo, 1979). We can see that the meridional wind stress expressed by Eq.(3′) is limited near the equator. In the north of the equator, there exists northerly wind stress ($Y < 0$) and in the south of the equator, it is the southerly wind stress ($Y > 0$). Obviously, Eq.(3′) represents the meridional wind stress which converges about the equator. The analysis of the observational data revealed in the first part of this study (Zhang et al., 2001) shows that the convergent center of the anomalous meridional wind stress is located a little north of the equator, not exactly along the equator as expressed in Eq.(3′). For simplicity, we take Eq.(3′) as the wind stress. Anyway, the form of the meridional wind stress taken here in Eq.(3′) mainly resembles the observations prior to the onset of El Niño.

By expanding variables in Eqs.(3) in terms of the parabolic cylinder function, we obtain the resolution of Eqs.(3) as below:

$$u = -\frac{1}{2\varepsilon^2 + 3} y^2 e^{-\frac{1}{4}y^2} \ , \tag{4a}$$

$$v = -\frac{2\varepsilon}{2\varepsilon^2 + 3} y e^{-\frac{1}{4}y^2} \ , \tag{4b}$$

$$h = \frac{1}{2\varepsilon^2 + 3}(2 - y^2) e^{-\frac{1}{4}y^2} \ . \tag{4c}$$

3.2 *Analyses of the solution*

Eqs.(4) show the solution of Eqs.(3) in response to the convergent meridional wind stress as expressed in Eq.(3′). We can see that the meridional oceanic currents v are anti–symmetric about the equator. It is southward to the north of the equator and northward to the south of the equator. The meridional current converges about the equator and, obviously, is the direct response to the convergent wind stress. The oceanic zonal currents u are symmetric about the equator. They flow westward in the region both to the south and to the north of the equator.

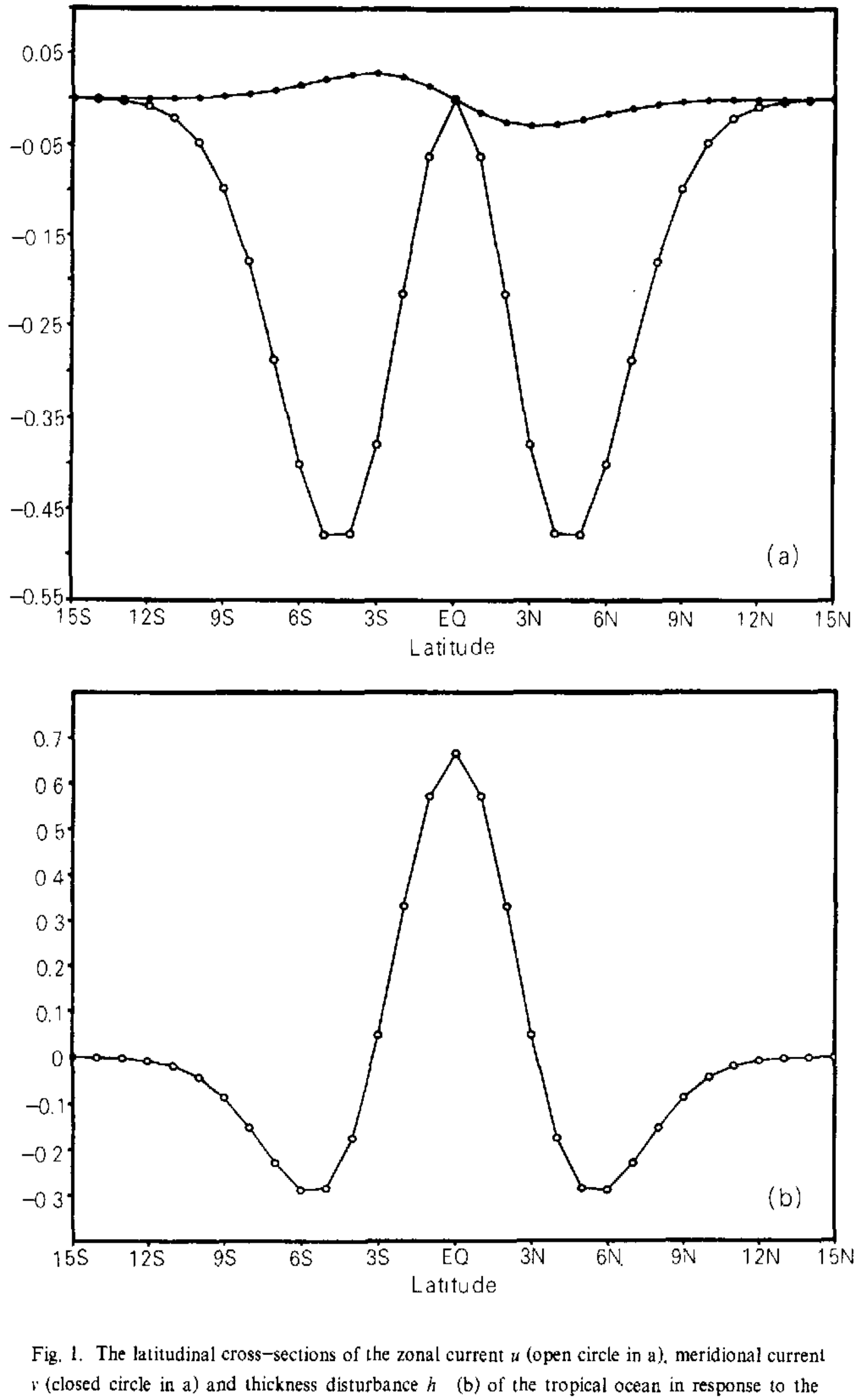

Fig. 1. The latitudinal cross-sections of the zonal current u (open circle in a), meridional current v (closed circle in a) and thickness disturbance h (b) of the tropical ocean in response to the meridional wind stress converging about the equator.

and vanishes at the equator. The zonal currents are the reflection of the Ekman drift of the meridional oceanic currents. The thickness disturbance h is symmetric about the equator too. In the vicinity of the equator are the positive thickness disturbances with the largest at the equator. Away from the equator, i.e., to the north of $y=2^{1/2}$ and to the south of $y=-2^{1/2}$, are the negative thickness disturbances. By taking the parameters $\beta=2.28\times10^{-11}s^{-1}m^{-1}$, $\varepsilon=0.05$ and $c=2.89$ m/s, we show u, v and h in the tropical ocean in Fig. 1.

The physical reasons for the distributions of u,v and h shown in Fig. 1 are apparent. If the meridional wind stress which converges about the equator is exerted on the tropical ocean, the sea water driven by the converging wind stress moves toward the equator and piles up there. The downwelling near the equator and upwelling away from the equator lead to the mixed layer near the equator becoming thicker and that away from the equator thinner. Furthermore, because of the Ekman drift of the meridional oceanic currents, the westward oceanic currents are formed on both sides of the equator, which transfer the surface sea water westward. Such distribution of the oceanic currents is shown in Fig. 2.

4. Non–stationary response of the equatorial ocean to convergent meridional wind stress

In the above section, the convergent meridional wind stress is considered to be evenly distributed in the zonal direction and acts on the whole zonal area. In fact, as revealed in the first part of this study (Zhang et al., 2001), observations show that the convergent area concentrates in a certain area in the eastern equatorial Pacific prior to the onset of El Niño events. Moreover, in the stationary response of the tropical ocean to the wind stress, there is no wave process and the evolution features of the oceanic response cannot be identified. Therefore, in this section we study the non–stationary response of the tropical ocean to the meridional wind stress converging about the equator in a limited zonal width.

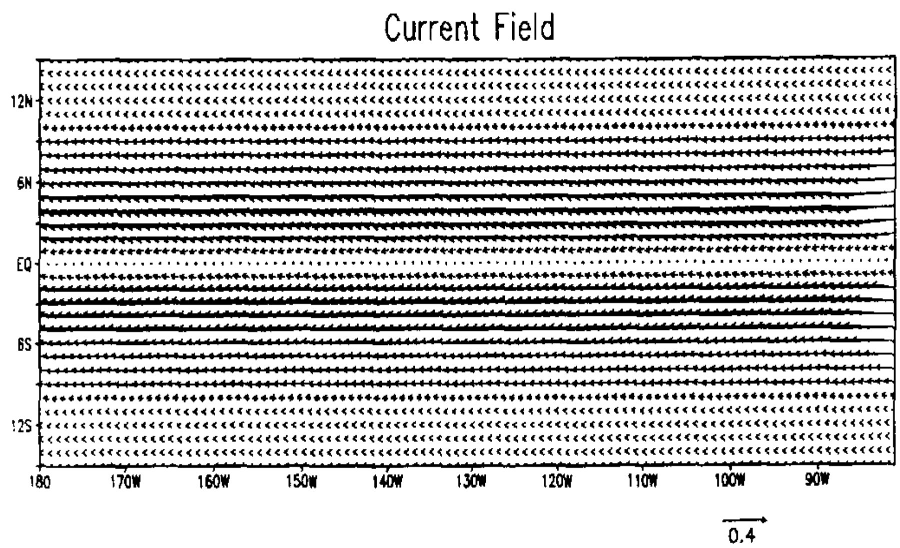

Fig. 2. The currents of the tropical ocean in response to the meridional wind stress converging about the equator.

4.1 *Simplification and resolution*

By introducing the long wave approximation (Gill, 1980) to Eqs.(2), we can write Eqs.(2) as

$$\frac{\partial u}{\partial t}-\frac{1}{2}yv+\frac{\partial h}{\partial x}+\varepsilon u=0\ , \tag{5a}$$

$$\frac{1}{2}yu+\frac{\partial h}{\partial y}=Y\ . \tag{5b}$$

$$\frac{\partial h}{\partial t}+\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\varepsilon h=0\ . \tag{5c}$$

According to Gill (1980), by introducing new variables

$$q=h+u\ , \tag{6a}$$

$$r=h-u\ , \tag{6b}$$

then we have

$$u=(q-r)/2\ , \tag{7a}$$
$$h=(q+r)/2\ . \tag{7b}$$

Substitute the new variables in Eqs.(6) into Eqs.(5) and then expand the variables q,r,v and Y by the parabolic cylinder function $D_m(y)$, i.e.,

$$(q,r,v,Y)(x,y,t)=\sum_{m=0}^{\infty}(q_m,r_m,v_m,Y_m)(x,t)D_m(y)\ . \tag{8}$$

For the meridional wind stress which converges about the equator, same as last section, we set $Y_m=0$ $(m\neq 1)$. Considering that the wind stress is confined in a limited zonal width, Y_1 can be taken as

$$Y_1=\begin{cases}-\cos kx\ , & |x|<L\\ 0\ , & |x|>L\end{cases} \tag{9}$$

and thus the meridional wind stress can be expressed as

$$Y=Y_1 D_1(y)\ , \tag{10}$$

where $k=\pi/2L$. Here we can see that in the forcing area the meridional wind stress Y in Eq.(10) converges about the equator with the strongest convergence at $x=0$ and is confined in a limited zonal width between the interval $[-L,L]$.

The solution of Eqs.(5) for the non-stationary response of the tropical ocean to the forcing of the convergent meridional wind stress expressed in terms of Eq.(10) is as follows:

$$u=\frac{q_2 D_2(y)-r_0 D_0(y)}{2}\ , \tag{11a}$$

$$v=v_1 D_1(y)\ , \tag{11b}$$

$$h = \frac{q_2 D_2(y) + r_0 D_0(y)}{2} \quad . \tag{11c}$$

The expressions of r_0, q_2 and v_1 are as below.

1) When $x > L$, i.e., to the east of the forcing area,

$$r_0 = q_2 = v_1 = 0 \quad . \tag{12a}$$

2) When $|x| < L$, i.e., within the forcing area,

$$q_2 = \frac{4k\varepsilon}{9\varepsilon^2 + k^2} e^{3\varepsilon(x-L)} - \frac{4k\varepsilon}{9\varepsilon^2 + k^2} \sin kx - \frac{6\varepsilon^2 + 2k^2}{9\varepsilon^2 + k^2} \cos kx$$
$$+ \frac{e^{-\varepsilon t}}{9\varepsilon^2 + k^2} \left\{ 4k\varepsilon \sin k(x + \frac{t}{3}) + (6\varepsilon^2 + 2k^2) \cos k(x + \frac{t}{3}) \right\} \left[H(\frac{t}{3}) - H(x - L + \frac{t}{3}) \right]$$
$$r_0 = 2q_2 - 2Y_1 = 2q_2 + 2\cos kx \quad ,$$
$$v_1 = \frac{4}{3}\frac{\partial q_2}{\partial x} + \frac{\partial}{3}\varepsilon Y_1 - \frac{\partial}{3}\frac{\partial Y_1}{\partial x} = \frac{4}{3}\frac{\partial q_2}{\partial x} - \frac{2}{3}\varepsilon \cos kx - \frac{2}{3} k \sin kx \quad . \tag{12b}$$

3) When $x < -L$, i.e., to the west of the forcing area,

$$q_2 = \frac{4k\varepsilon}{9\varepsilon^2 + k^2} \left\{ e^{3\varepsilon(x+L)} H(x + L + \frac{t}{3}) + e^{3\varepsilon(x-L)} H(x - L + \frac{t}{3}) \right\}$$
$$+ \frac{e^{-\varepsilon t}}{9\varepsilon^2 + k^2} \left[(2k^2 + 6\varepsilon^2) \cos(x + \frac{t}{3}) + 4k\varepsilon \sin k(x + \frac{t}{3}) \right]$$
$$\cdot \left[H(x + L + \frac{t}{3}) - H(x - L + \frac{t}{3}) \right] ,$$
$$r_0 = 2q_2 \quad ,$$
$$v_1 = \frac{4}{3}\frac{\partial q_2}{\partial x} \quad . \tag{12c}$$

In the above expressions, $H(x)$ is the Heaviside function.

4.2 *Solution analysis*

According to the characteristics of the parabolic cylinder function $D_m(y)$ (see Wang and Guo, 1979), from the solutions (11) we can see that the meridional oceanic current v is anti–symmetric about the equator. The zonal current u and the thickness disturbance h are symmetric about the equator. So the symmetry of the solutions is the same as that in the stationary response discussed in the last section.

By examining Eqs.(11) and (12), we can find that the solutions of the oceanic response mainly depend on the form of q_2 except within the forcing area where there also exists the effect of the wind stress Y_1. Therefore, the evolution of the oceanic disturbances under the forcing of the convergent meridional wind stress can be discussed through analyzing the expression of q_2.

From the expression (12a), we can see that in the east of the forcing area ($x > L$), r_0, q_2 and v_1 all vanish. The ocean is not affected by the wind stress and there are no oceanic disturbances in this area. This is simply because the meridional wind stress cannot excite the eastward propagating Kelvin wave in the ocean, so that no disturbances exist in the east of the forcing area.

In the forcing area ($|x| < L$), q_2 in (12b) consists of both the stationary and the non–stationary parts. The non–stationary part represents the westward propagating Rossby wave with the width of $2L$. Beginning at $t=0$ when the forcing exerts, the Rossby wave starts to move westward and finally moves out of the forcing area into the west of the forcing area. At time t_0, the area with the width $[-L, L-t_0/3]$ in the force area is affected by the Rossby wave, while in the interval $[L-t_0/3, L]$ it becomes stationary. When $t>6L$, the Rossby wave completely moves out of the forcing area and the non–stationary part does not exist. In the forcing area the oceanic disturbances become stationary and q_2 can be expressed as

$$q_2 = \frac{4k\varepsilon}{9\varepsilon^2 + k^2} e^{3\varepsilon(x-L)} - \frac{4k\varepsilon}{9\varepsilon^2 + k^2} \sin kx - \frac{6\varepsilon^2 + 2k^2}{9\varepsilon^2 + k^2} \cos kx \ . \tag{13}$$

In the west of the forcing area ($x<-L$), the solutions are shown in (12c). At time t, the last part of q_2 moves westward at the non–dimensional speed $1/3$ and exists in the area with the width $2L$ at the interval $[-L-t/3, L-t/3]$. Obviously, it is the Rossby wave excited by convergent meridional wind stress in the forcing area. Because of the dissipation, its amplitude reduces by the factor $\exp(-\varepsilon t)$. In the interval $[-L-t/3, L-t/3]$, the q_2 can be expressed as

$$\begin{aligned} q_2 = &\frac{4k\varepsilon}{9\varepsilon^2 + k^2} e^{3\varepsilon(x+L)} \\ &+ \frac{e^{-\varepsilon t}}{9\varepsilon^2 + k^2}\left[(6\varepsilon^2 + 2k^2)\cos k\left(x+\frac{t}{3}\right) + 4k\varepsilon \sin k\left(x+\frac{t}{3}\right)\right] . \end{aligned} \tag{14}$$

After $t=6L$ when the Rossby wave moves out of the forcing area, q_2 becomes stationary and exists in the west of the forcing area at the interval $[L-t/3, -L]$. It can be expressed as

$$q_2 = \frac{4k\varepsilon}{9\varepsilon^2 + k^2}\left[e^{3\varepsilon(x+L)} + e^{3\varepsilon(x-L)}\right] . \tag{15}$$

By taking non–dimensional width of the forcing area to be $2L=8.8$, which corresponds to dimensional width of about 20 degrees longitude, Fig. 3 shows the distributions of the oceanic disturbances at non–dimensional time 9 (Fig. 3a), 27 (Fig. 3b), 54 (Fig. 3c) and 81 (Fig. 3d), respectively. We can see clearly from Fig. 3 that the Rossby wave is excited by the convergent meridional wind stress in the forcing area. It then moves out of the forcing area and keeps on propagating westward as the time increases. In Fig. 3a, the Rossby wave has not moved out of the forcing area completely and within the forcing area the stationary response has not been reached. In Fig. 3b, the Rossby wave has almost moved out of the forcing area and it becomes stationary in the most part of the forcing area. In Fig. 3c and Fig. 3d, it is obvious that there exist three different kinds of the disturbance patterns distributed zonally. Within the forcing area in the interval $[-L, L]$, there are two negative centers of the thickness disturbance h on both sides of the equator. They are two lows and are symmetric about the equator. Accompanying the two lows there appear cyclones of oceanic current fields. In the west of the forcing area, there are two parts of the disturbance fields. Within the interval $[-L-t/3, L-t/3]$, it is the non–stationary Rossby wave which moves westward continuously. Corresponding to the Rossby wave, there are two positive centers of the thickness disturbance h on both sides of the equator. They are two highs and are symmetric about the equator. Accompanying the two highs, anticyclones of oceanic current fields appear. Within the interval $[L-t/3, -L]$ between the Rossby wave and the forcing area, the disturbances are stationary. At the equator and on both sides within about 3° latitude, there

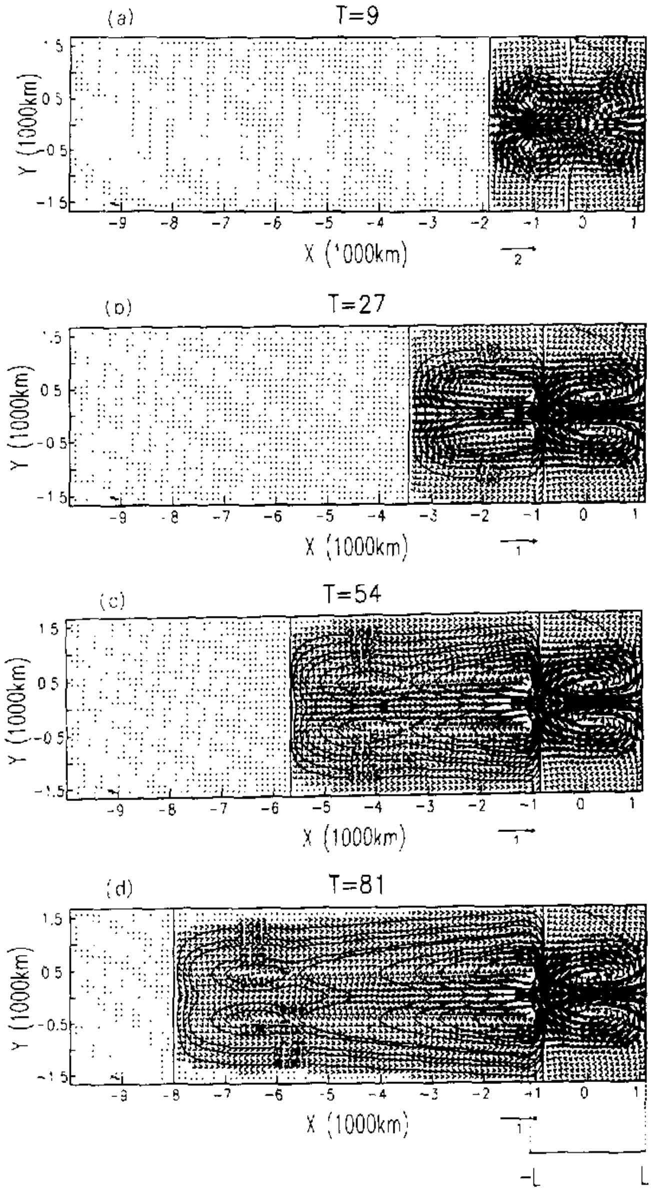

Fig. 3. The distribution of the oceanic disturbances at the non-dimensional time $t = 9$ (a), $t = 27$ (b), $t = 54$ (c) and $t = 81$ (d). Arrows are the oceanic currents. Real and dotted lines are the positive and negative thickness disturbances, respectively.

exist strong oceanic currents which flow westward. On both sides of the equator, the positive thickness disturbance h exists. Because of the effect of dissipation, the amplitudes of the disturbances attenuate gradually as the disturbances move westward.

Figure 4 shows the time–longitudinal cross–sections of the thickness disturbance h (a) and zonal current u (b) along the equator, respectively. From Fig. 4 we can see that because of the westward propagation of the Rossby wave, both h and u expand westward along the equator as the time increases. The evolution of h in Fig. 4a shows that the Rossby wave

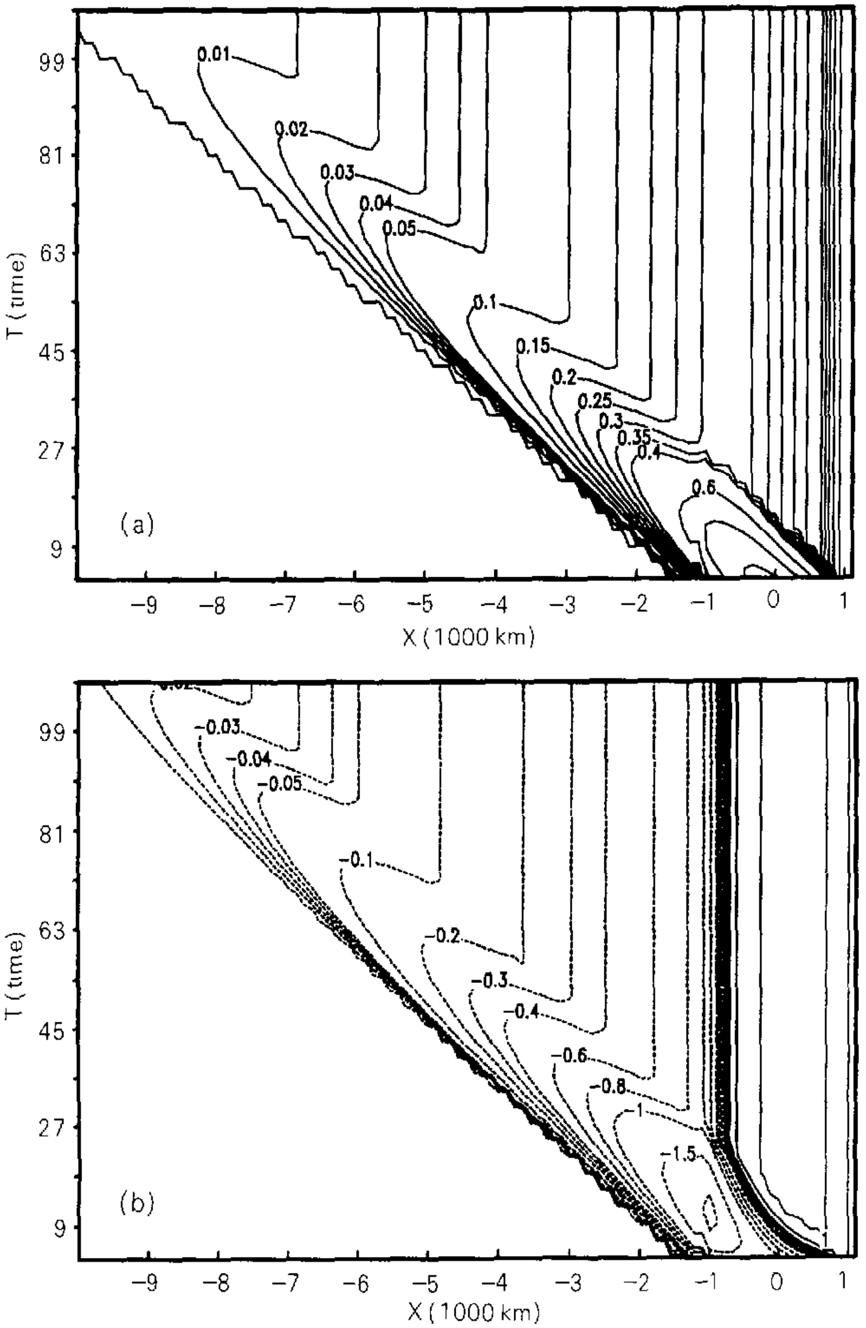

Fig. 4. Time−longitudinal cross−sections of the thickness disturbance h (a) and zonal current u (b) along the equator in the oceanic mixed layer. The real and dotted lines represent the positive and negative disturbances, respectively.

makes the oceanic mixed layer become thicker along the equator. Due to the effect of the dissipation, the amplitude of the positive thickness disturbances gets smaller gradually as the disturbances move westward. Therefore, the oceanic mixed layer becomes thicker in the west than in the east. In Fig. 4b, we can see that there exist oceanic currents which flow westward along the equator. As the time increases, the scope of the westward currents expands to the west because of the westward propagation of the Rossby wave. Also due to the effect of dissipation, the amplitude of the oceanic currents weakens gradually as the oceanic currents expand to the west.

5. Summary and discussion

In this paper, based on the observations (Zhang et al., 2001), an ideal meridional wind stress which converges about the equator is set up. By utilizing a shallow water model in the tropical ocean, we investigate the dynamical processes of the tropical ocean in response to the wind stress. The main conclusions are as follows.

For the stationary oceanic response when the forcing of the convergent meridional wind stress is zonally distributed evenly, the meridional oceanic currents on both sides of the equator flow toward the equator. The piling up of the sea water near the equator makes the oceanic mixed layer near the equator become thicker. Correspondingly, the westward oceanic currents exist near the equator.

For the non–stationary oceanic response when the forcing of the convergent meridional wind stress is limited within a certain zonal width, the westward propagating Rossby wave symmetric about the equator is excited. In the west of the forcing area, the Rossby wave makes the oceanic mixed layer become thicker. And the westward oceanic currents appear near the equator. The scope of the thickened mixed layer and westward oceanic currents expands westward as the time increases because of the westward propagation of the Rossby wave. Due to the dissipation, the amplitudes of the positive thickness disturbances and the westward currents attenuate westwards.

The analyses of the observational data in the first part of this study (Zhang et al., 2001) reveal that prior to the increasing of SSTA in the eastern equatorial Pacific by more than half a year, there appears convergence of the anomalous meridional wind stress in the eastern equatorial Pacific. In this paper, we can see that when the meridional wind stress converging about the equator exerts on the tropical ocean, the oceanic Rossby wave can be excited. The westward propagating Rossby wave makes the oceanic mixed layer near the equator become thicker and the most prominently thickened region is near the forcing area due to the effect of dissipation. Especially when the forcing of the convergent meridional wind stress is within the eastern equatorial Pacific, the climatological mean state of the thin mixed layer there can be thickened, which is favorable for the El Niño occurring. On the other hand, the westward oceanic currents in the vicinity of the equator can transfer the sea water in the mixed layer from the eastern tropical Pacific to the west. According to the theory of Wyrtki (1975), the westward oceanic currents near the equator are favorable for the upper layer sea water to pile up in the western tropical Pacific and thus to accumulate the energy for the following warm event. Here we can see that prior to an El Niño, the sea water in the upper layer tropical Pacific can be driven to the west not only by the easterly wind stress anomalies associated with the strengthening of the trade winds as proposed by Wyitki (1975), but also by the convergent meridional wind stress anomalies which appear in the eastern equatorial Pacific.

In the past, the researches emphasized the importance of the zonal wind stress anomalies in an El Niño event. Compared to the zonal anomalies, the data analysis in the first part of our present study (Zhang et al., 2001) shows that the convergent meridional wind stress anomalies appearing in the eastern equatorial Pacific prior to an El Niño have a comparable contribution to the increasing of the SSTA in the eastern equatorial Pacific. The convergence of the meridional wind stress anomalies can lead to an El Niño through increasing the mixed layer depth and transporting more sea water to the west to accumulate the energy for the following warm event. The westerly anomalies can induce eastward propagating downwelling Kelvin waves which are also favorable for the subsequent warming in the central and eastern Pacific (e.g., Wyrtki, 1975; McCreary, 1976; Busalacchi and O'Brien, 1981; Zhang and Huang, 1998; Huang et al., 1998). Our results here suggest that both the westerly anomalies in the western tropical Pacific and the convergence of the anomalous meridional wind stress in the eastern equatorial Pacific play an important role in the occurrence of El Niño events.

In the present study, a simple tropical oceanic model is used to investigate the response of the tropical ocean to the converging wind stress. For fully understanding the dynamics and thermodynamics of the response of the tropical ocean to the converging wind stress, it is necessary to use a more complicated oceanic model. In fact, the data analysis in the first part of this study (Zhang et al., 2001) shows that besides the converging meridional wind stress anomalies in the eastern equatorial Pacific, another significant correlation area appears in the tropical western Pacific to the northeast of Australia. The role of the leading meridional wind anomalies in this area in the occurrence of El Niño events is worth to be studied in the future investigation.

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热带太平洋经向风应力异常与 El Niño 的发生 ——(II)动力学分析

张人禾　　赵　刚

摘　要

第一部分(Zhang et al., 2001)的资料分析表明，El Niño 事件发生之前在赤道中东太平洋存在着显著的异常经向风应力辐合。为了分析这种超前的辐合经向风应力距平在其后的 El Niño 事件发生中的动力作用，本文利用简单热带海洋动力学模式，从动力学上研究了热带海洋对关于赤道辐合的理想经向风应力强迫的响应，指出赤道东太平洋出现在 El Niño 事件之前的辐合经向风应力异常有利于 El Niño 事件的发生。辐合的经向风应力强迫作用于热带海洋，会激发出西传的 Rossby 波，使得赤道附近的海洋混合层变厚。由于耗散的影响，最大的增厚区位于强迫区域。当这个强迫作用于赤道东太平洋时，这将有利于 El Niño 事件发生；另一方面，Rossby 波响应在赤道及其附近使得表层海水向西流动，中东太平洋表层水的不断向西输送有利于表层水在西太平洋堆积，为后来暖事件的发生累积能量。

关键词：经向风应力，厄尔尼诺(El Niño)