The Response of Equatorial Oceans to a Relaxation of the Trade Winds

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ABSTRACT

The trade winds over the central Pacific are observed to weaken several months after the appearance of anomalously warm surface waters in the eastern equatorial Pacific Ocean. The following results obtained with a numerical model indicate how this relaxation of the winds affect the later stages of El Niño. A weakening of the westward trade winds causes a zonal redistribution of heat in the equatorial oceans and a warming of the eastern part of the basin. The warming depends on the zonal extent of the region over which the winds relax, and on the length of time T for which the winds relax. As T increases the warming in the east increases until it asymtotes to a maximum value when T exceeds the adjustment time of the basin (which is ~400 days in the case of the Pacific Ocean). Maximum heating is associated with a permanent weakening of the winds, unless the winds reverse direction and become eastward. Even weak eastward winds for a short period can cause disproportionately large temperature increases (because of nonlinear mechanisms).

In the region where the winds relax, the heating is due to convergence of surface waters on the equator, and advection by accelerating eastward surface currents. As the time scale T increases, the acceleration becomes less pronounced. East of the region where the winds relax, Kelvin waves suppress the thermocline but leave the sea surface temperature unchanged in linear models. In nonlinear models advection by eastward currents in the wake of Kelvin waves can cause a warming, even at the surface. For winds with a realistic spatial and temporal structure the identification of these waves is difficult.

1. Introduction

The occurrence of El Niño events (during which anomalously warm surface waters appear in the eastern equatorial Pacific Ocean) is correlated with a weakening of the trade winds (Wyrtki, 1975, 1977, 1979). The weakening occurs several months after the appearance of warm surface waters off the coast of Peru (Rasmussen and Carpenter, 1981) and can therefore not be responsible for the initiation of El Niño events. The oceanic response to the change in the wind stress, nonetheless, is of considerable interest because it will affect the later stages of El Niño. For example, the weakening of the winds could sustain El Niño events if the following argument is correct.

Under normal conditions the westward surface winds over the Pacific cause the thermocline to slope downward from east to west as shown in Fig. 1. The sea surface temperatures and heat content of the ocean therefore are smaller in the east than in the west. When the winds relax, the thermocline reverts to a horizontal position so that heat is transferred from west to east. This contributes to the abnormally high sea surface temperatures observed in the eastern Pacific during El Niño.

The manner in which heat is redistributed in the ocean depends entirely on the manner in which the winds relax. Two idealized cases have been studied. In the one case it is assumed that the relaxation of the wind is sudden. In the other case it is assumed that the winds weaken gradually on a time-scale long compared to the adjustment-time of the ocean.

McCreary (1976) and Hurlburt *et al.* (1976) solve the shallow-water equations to determine how an abrupt weakening of westward winds affects the ocean. When the winds die out, the wind-driven westward current in these models decelerates, and then becomes an accelerating eastward equatorial jet. The acceleration is caused by the eastward pressure force, which had been maintained by the westward winds, and which is unbalanced when the winds relax. (This pressure force can be inferred from the zonal density gradients in Fig. 1.) The eastward equatorial jet is convergent and causes a suppression of the thermocline, especially in the east, so that the heat content of the ocean increases. An equatorial Kelvin wave that emanates from the western coast eliminates the eastward pressure force, arrests the acceleration of the eastward jet, and stops the warming of the eastern part of the basin. (In linear models the warming is not associated with higher sea surface temperatures, but with an increase in the depth of the thermocline. See Section 3.)

The one and two-level models used by McCreary (1976) and Hurlburt *et al.* (1976) have the virtue of

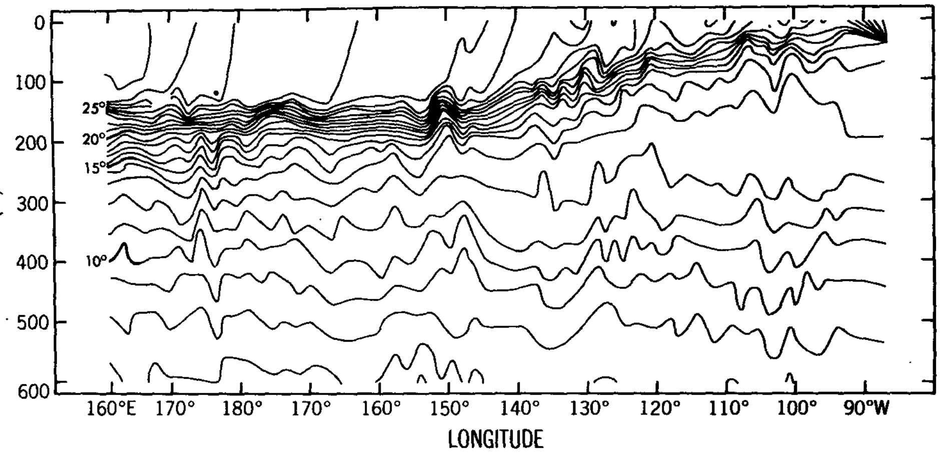

FIG. 1. Temperature as a function of longitude and depth along the equator in the Pacific Ocean as measured by Colin *et al.* (1971).

simplicity (so that the oceanic response can readily be calculated and analyzed), but these models also have severe limitations. For example, when the model-oceans are in equilibrium with a uniform zonal wind stress τ^x, they are in a state of rest because a zonal pressure gradient P_x exactly balances τ^x

$$\tau^x/h = P_x/\rho. \quad (1)$$

(The pressure gradient is related to the slope of the thermocline which is at a mean depth h). In reality, steady winds maintain a pressure gradient *and* a current system (which includes the Equatorial Undercurrent). The simplest model capable of simulating such currents is that the of Cane (1979). His results (Cane, 1980) confirm that an abrupt relaxation of westward winds gives rise to an accelerating eastward surface jet, but the model shows that it is in effect the Equatorial Undercurrent which surfaces, broadens and intensifies. A Kelvin wave from the western coast again stops the warming in the east, and reflects off the eastern coast as an equatorially trapped Rossby wave. The equatorial ocean has essentially adjusted to the new wind conditions in the wake of this Rossby wave. The adjustment time of the equatorial ocean is therefore equal to the time it takes a Kelvin wave to propagate eastward plus the time it takes a Rossby wave to propagate westward across the basin. In the case of the Atlantic Ocean this is estimated to be 150 days, in the case of the Pacific Ocean, 450 days. These values increase rapidly with latitude.

Studies of the adjustment of the ocean to an abrupt change in the intensity of the winds are of direct relevance to the Indian Ocean. Over the Atlantic and Pacific Oceans, however, large-scale wind variances occur over a period of several months, not days or weeks. (See for example Wyrtki's (1975) description of wind fluctuations during the 1972 El Niño.) The following example demonstrates that the results for an abrupt change in the intensity of the wind may not be directly relevant to situations in which the winds weaken gradually. Suppose that the winds fluctuate on a time-scale long compared to the adjustment time of the ocean. The ocean, in such a case, is always in an adjusted state and at each moment is in equilibrium with the winds at that moment. Eq. (1) is always valid so that the zonal slope of the thermocline is proportional to the intensity of the wind at all times. At these low frequencies there is essentially no phase lag between the forcing and the response because the ocean has no "memory" of past winds (Philander, 1979a). There is evidence that the low frequency variability of the equatorial oceans does indeed correspond to such an equilibrium response. Measurements show that the zonal pressure gradient, and the zonal component of the wind stress vary almost in phase on seasonal time scales in the Atlantic Ocean (Katz and collaborators, 1977) and on interannual time scales in the Pacific Ocean (Barnett, 1977). The tidal records analyzed by Enfield and Allen (1980) also confirm an equilibrium response at low frequencies in low latitudes. They find that sea level fluctuations at different points along the eastern coast of the equatorial Pacific Ocean all have the same phase. (Phase differences appear only outside the tropics.)

Gradually and abruptly changing winds are seen to affect the ocean in completely different ways. When the winds change suddenly there are propagating wave fronts and accelerating currents, which redistribute heat zonally. When the winds vary slowly, the transfer of heat is imperceptible because it is associated with very small velocity differences. Neither of these two cases describes the fluctuations of winds over the Pacific Ocean accurately. On the one hand, the winds do not weaken instantaneously. On the other hand, the fluctuations are not confined

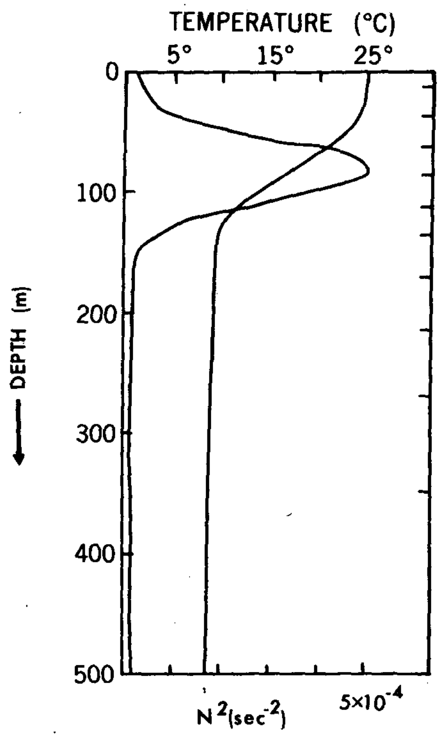

FIG. 2. The initial temperature and Brunt-Väisälä frequency in the model. The positions of grid points are shown along the ordinate. Below 500 m the temperature decreases linearly to 4°C at a depth of 3000 m.

to very low frequencies only. A study of the oceanic response to winds that vary gradually, but on time scales shorter than the adjustment time of the ocean, will therefore be valuable. It will be particularly interesting to know whether propagating waves are evident. The data analyzed by Wyrtki (1975, 1977), reveal interesting phase differences between sea level changes at different locations in the Pacific Ocean. But the data are inadequate to determine whether these phase differences are associated with propagating waves or with variable currents or whether they have some other explanation. The calculations to be described in this paper will shed light on this matter.

Section 2 of this paper contains a description of a numerical model which is used to determine how the ocean responds to (Section 3) variations in the time over which the winds relax permanently, (Section 4) variations in the length of time for which the winds relax temporarily, and (Section 5) the appearance of eastward winds. The results are summarized and discussed in Section 6.

2. The model

The equations of motion (the primitive equations) are simplified by making the Boussinesq and hydrostatic approximations and by assuming an equation of state of the form $\rho = \rho_0(1 - \alpha T)$, where ρ is the density, T the temperature, $\alpha = 0.0002°\mathrm{C}^{-1}$ is the coefficient of thermal expansion and $\rho_0 = 1\ \mathrm{g\ cm^{-3}}$. The coefficients of horizontal (ν_H) and vertical (ν_v) eddy viscosity, and the thermal diffusivity (κ) are assumed to have the constant values

$$\nu_v = 10\ \mathrm{cm^2\ s^{-1}}; \quad \nu_H = 2 \times 10^7\ \mathrm{cm^2\ s^{-1}};$$

$$\kappa_v = 1\ \mathrm{cm^2\ s^{-1}}; K_H = 10^7\ \mathrm{cm^2\ s^{-1}}.$$

Philander and Pacanowski (1980a) discuss the sensitivity of the model to the values of these parameters. The equations are solved numerically by using the method described by Bryan (1969) who discusses the finite differencing schemes in detail.

The model ocean is a rectangular box with a longitudinal extent of 4800 km, a latitudinal extent of 2800 km (with the equator in the center) and a depth of 3000 m. In a horizontal plane the 70 × 70 grid points are spaced at regular intervals (of 40 and 79 km in the latitudinal and longitudinal directions, respectively). In the vertical the 16 grid points are spaced irregularly. Fig. 2 shows the distribution of grid points in the upper 500 m.

Motion is forced at the ocean surface by imposing a wind stress. The heat flux is zero.

$$\nu U_z = \tau^x; \quad \nu V_z = 0; \quad w = 0, \quad T_z = 0.$$

At the vertical walls each of the velocity components, and the heat flux is zero. The ocean floor is stress-free and has a temperature of 4°C.

The initial temperature (and Brunt-Väisälä frequency) is shown in Fig. 2. Since ours is a diffusive model, the temperature will immediately start to evolve to a linear profile even in the absence of any forcing. This diffusive change, with which no motion is associated, is a very slow one so that the thermocline only disappears on a time scale that is long compared to a decade. Here we concern ourselves with transient phenomena with much shorter time scales. We do not address questions concerning the maintenance of the thermocline, but confine our attention to the manner in which it is deformed by different forcing functions.

A uniform westward wind of intensity 0.5 dyn cm^{-2} is suddenly turned on initially (day zero) over a motionless ocean, and prevails for 300 days. By that time the equatorial ocean is in a state of equilibrium (Philander and Pacanowski, 1980a). The adjustment of the upper ocean to the suddenly imposed surface winds is effected by vertical modes trapped in and above the sharp shallow thermocline because of internal reflection there. This is a direct consequence of the shape of the Brunt-Väisälä frequency (Fig. 2) which has effectively two discontinuities (see Philander and Pacanowski, 1980a). These modes have an equivalent depth of ~25 cm.

In this paper we start our calculations at day 300 and study the oceanic response to changes in the surface winds. Oceanographic conditions at day 300 are shown in Figs. 3a and 3b and Fig. 4a. Note that

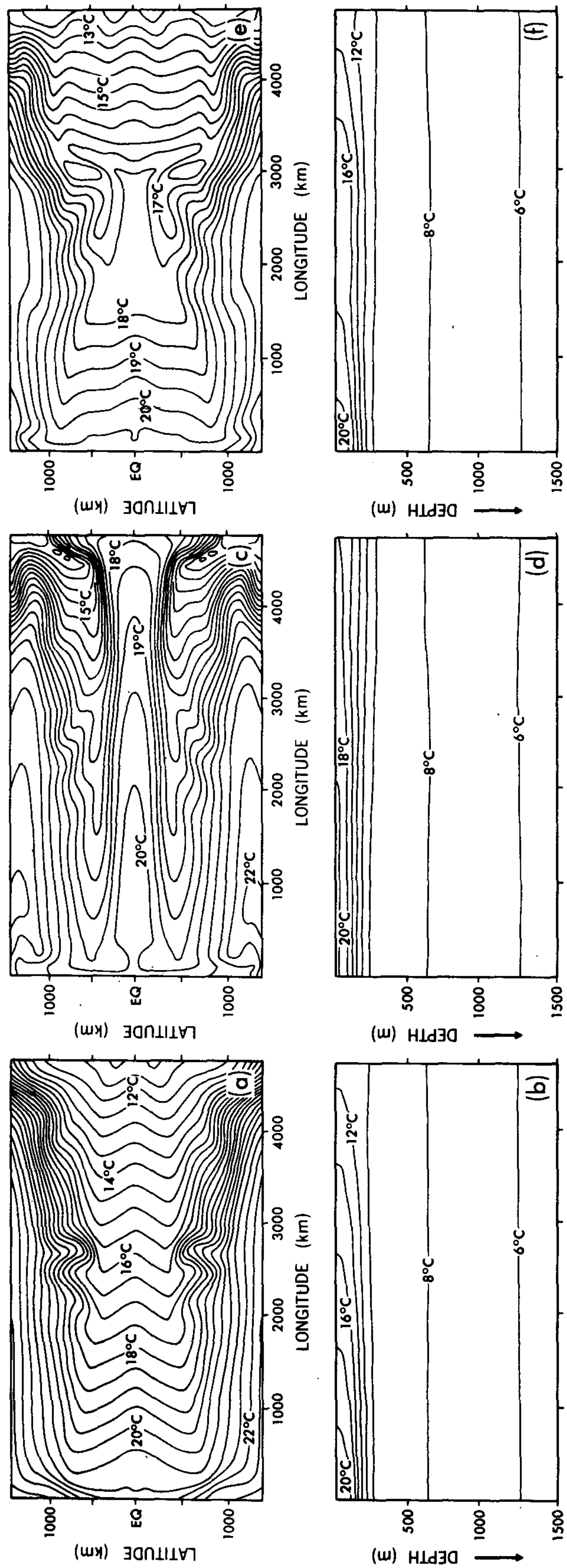

FIG. 3. Isotherms at a depth of 12.5 m at (a) the instant when the uniform westward winds are turned off, (c) 100 days after the winds had been turned off everywhere, and (e) 100 days after the winds had been turned off between meridians A and B. Isotherms in the equatorial plane, at corresponding times, are shown below. Figs. 3b, 3d and 3f.

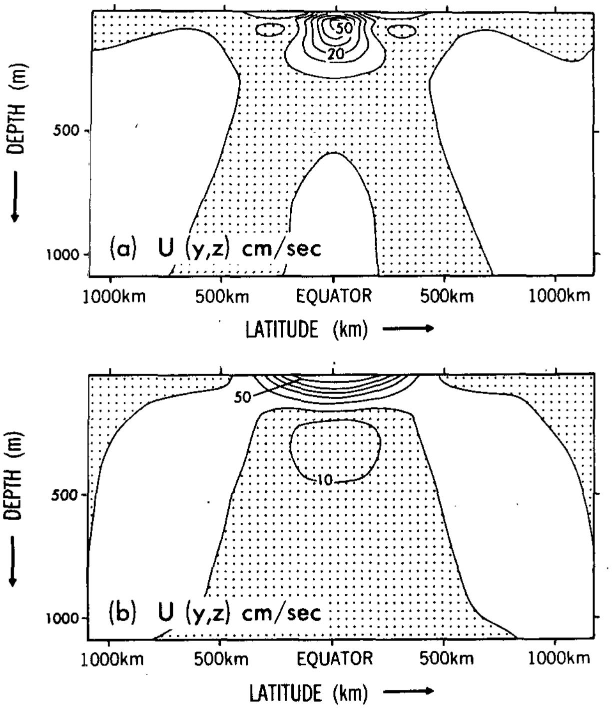

FIG. 4. Zonal currents across the central meridian of the basin (a) at the instant when the uniform winds are turned off and (b) 50 days after the winds had been turned off everywhere. Motion is westward in shaded areas.

the stratification has spatial variations so that the vertical profile of the Brünt-Väisälä frequency N varies considerably with position. It is therefore remarkable that the earlier mentioned thermocline trapped modes, which are responsible for the adjustment of the upper ocean, are not significantly affected by the nonuniformity of N. This will be evident from the results presented in the next section.

3. Variations in the manner in which the winds relax

a. *Abrupt relaxation of uniform winds*

In the introduction we mentioned several studies of the response of the ocean to an abrupt relaxation of the winds. The models used in those studies have one (or two) layers. Each layer has a constant temperature. To infer temperature variations it therefore is necessary to postulate a relation between the thermocline depth and the sea surface temperature. This is a difficult problem: in some parts of the ocean—the western equatorial Atlantic, for example, considerable vertical movements of the thermocline leave the sea surface temperature almost unaffected. In other parts of the ocean—the eastern equatorial Atlantic, for example, slight vertical displacements of the thermocline have a large effect on the sea surface temperature (see Merle, 1980). Our multilevel model overcomes this difficulty, and permits us to be quantitative about temperature changes, because temperature is a variable in our model. We therefore start by reinvestigating the response of the ocean to a weakening of the winds.

In our first experiment, uniform westward winds which had prevailed for 300 days, are abruptly and permanently turned off. Initially the zonal currents have the structure shown in Fig. 4a. Note that the eastward Equatorial Undercurrent, which extends to the surface, has a subsurface core. The initial temperature field (Figs. 3a and 3b) implies the existence of an eastward pressure force in the upper ocean. This pressure force is balanced by the wind stress close to the surface, and is balanced by diffusion at greater depths. When the wind suddenly stops blowing the pressure force is unbalanced in the surface layers and therefore causes eastward acceleration there (Fig. 5a). Conditions below the surface layers (Fig. 5b) are at first almost unaffected because diffusion acts gradually. The accelerating equatorial jet advects a considerable amount of heat eastward so that the eastern side of the basin heats rapidly (Fig. 5c). This jet—its structure is shown in Fig. 4b—is associated with convergent motion and equatorial downwelling.

The initial acceleration of the surface flow is longitudinally uniform. Free modes of the ocean are therefore excited at the coasts to satisfy boundary conditions there. The vertical structure of these modes depends on the stratification of the ocean. The most important mode is an equatorial Kelvin wave excited at the western coast. It has a large amplitude in and above the thermocline only, and therefore affects oceanic conditions in the surface layers (where the accelerating jet is) *and* in the thermocline. The dashed line in Fig. 5 shows the propagation of this Kelvin wave front across the basin. It is seen to eliminate the zonal density gradient, to arrest the acceleration of the surface jet, to stop the warming of the upper ocean, and to initiate the deceleration of the Equatorial Undercurrent.

According to linear theory, Rossby wave fronts are excited initially at the eastern coast. In the nonlinear model, these waves apparently have too small an amplitude to propagate westward against the eastward jet. A Rossby wave is excited only by the reflection of the Kelvin wave off the eastern coast. The ocean rapidly returns to a state of rest in the wake of this Rossby wave.

The relaxation of the winds causes temperatures to increase by as much as 5°C in the eastern part of the basin. This is accomplished by means of advection which redistributes heat zonally so that the thermocline reverts to a horizontal position (Figs. 3c and 3d). At one stage, the deeper isotherms actually slope downwards to the east. See the 10°C

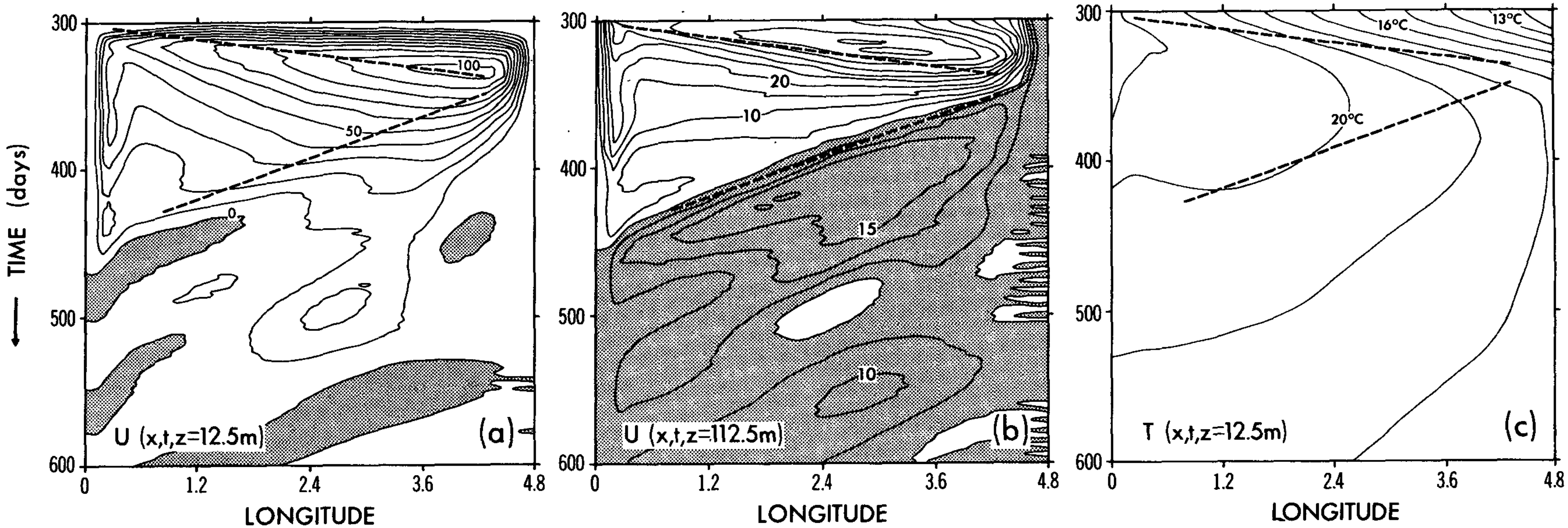

FIG. 5. The evolution of conditions along the equator after the abrupt relaxation of uniform westward winds. The dashed lines indicate the eastward propagating Kelvin, and westward propagating gravest equatorially trapped Rossby modes, with equivalent depths of 25 cm. (a) Zonal currents at 12.5 m, (b) zonal currents at 112.5 m, and (c) temperature at 12.5 m. Motion is westward in shaded regions. Units for the velocities are cm s^{-1}, for longitude 10^3 km.

isotherm in Fig. 3d, for example. The associated westward pressure force is presumably the source of momentum for the westward subsurface flow in Fig. 5b.

b. Abrupt relaxation of spatially varying winds

In our second experiment the wind relaxes only over a part of the basin, between meridians A and B, say. Cane and Sarachik (1976) and McCreary (1980) studied the linear response of a one-level ocean to spatially varying winds. We summarize their results briefly. In the forced region (between meridians A and B) currents evolve as if that were a bounded region because waves excited at A and B are similar to waves excited at coasts. To the east of B, motion is due to Kelvin waves excited at A and B, plus Rossby waves due to the reflection of these Kelvin waves off the eastern coast. The net effect of all these waves is to alter the depth of the thermocline by an amount which (i) is independent of longitude and (ii) is proportional to the distance between A and B. The waves introduce only transient currents east of B, unless friction is important. Changes in conditions to the west of A are similar to changes to the east of B but are due to Rossby waves excited at A and B, plus their reflections off the western coast.

In a linear multilevel model the results are similar to those described above. In the unforced region waves modify the vertical structure of the temperature field (by raising or lowering the depth of the thermocline, for example) but the surface temperature remains unchanged. We shall now show that in a nonlinear model the sea surface temperature in the unforced region can change, because of advection.

In the nonlinear model the uniform winds, which had prevailed for 300 days, abruptly stop blowing between meridians A and B which are 1600 km and 3200 km from the western coast, respectively. Outside this band of meridians the winds are unchanged. Hence, after day 300

$$H(x) = 0.5\left[\tanh\left(\frac{x - 1600\ \mathrm{km}}{120\ \mathrm{km}}\right) - \tanh\left(\frac{x - 3200\ \mathrm{km}}{120\ \mathrm{km}}\right)\right].$$

The oceanic response is shown in Figs. 3e and 3f and 6a, 6b and 6c.

West of A there is a serious discrepancy between linear and nonlinear theory. Conditions there are practically unaffected by the change in wind conditions east of A. (The gradual change in temperature west of A in Fig. 6a is consistent with a diffusive cooling.) The reason for this is presumably related to the presence of the Equatorial Undercurrent which has a maximum speed of 55 cm s^{-1} in the region west of A.

Between meridians A and B the response is qualitatively similar to the response described in Section 3a. Amplitudes are smaller, however, the surface flow attains a maximum speed of only 60 cm s^{-1}, and the warming of the upper ocean is relatively small. The reason for this is the smaller amount of potential energy being released when the winds relax. (This energy is stored in the zonal density gradient.)

To the east of meridian B changes are qualitatively as predicted by linear theory. A Kelvin wave front from B causes an estimated acceleration of surface and subsurface flow and lowers the thermocline. Advection by currents in the wake of this wave increases the sea surface temperature. A Kelvin wave front from meridian A stops the acceleration and warming. Rossby waves due to reflections at the eastern coast are not clearly evident in Figs. 6a–6c, presumably because of the presence of an intense Equatorial Undercurrent. The intensified currents nonetheless weaken again, and after ~100 days return to their original state. The temperature to the east of B, however, remains higher than it originally was. This increase in temperature is essentially independent of longitude.

Figs. 3e and 3f show the temperature changes due to a relaxation of the wind over a confined region in the middle of the basin. Whereas a local weakening of the winds (between A and B) eliminates longitudinal gradients, a nonlocal weakening leaves these gradients unchanged but, in the region east of B, causes a uniform increase in temperature.

c. A gradual relaxation of uniform winds

When uniform winds weaken instantaneously then the oceanic response is composed of two parts: a longitudinally uniform current that accelerates almost linearly with time, plus a nondispersive Kelvin wave front with a time (t) and longitude (x) dependence of the form $S(x - ct)$. (Here c is the phase speed and S a stepfunction). These two components are so different that it is an easy matter to identify each. The Kelvin wave, for example, is clearly discernible in Fig. 5.

When the relaxation of the wind is gradual then there is no advantage in describing the response as consisting of two parts (as we did above) because each part has a complicated time-dependence and the two parts are not readily separable. This is evident in Fig. 7 which shows the evolution of the surface currents when the winds decay gradually according to the expressions

$$\tau^x = \begin{cases} -0.5\{1 - \exp[-4(t/T - 1)^2\}, & t < T \\ 0, & t > T. \end{cases} \quad (2)$$

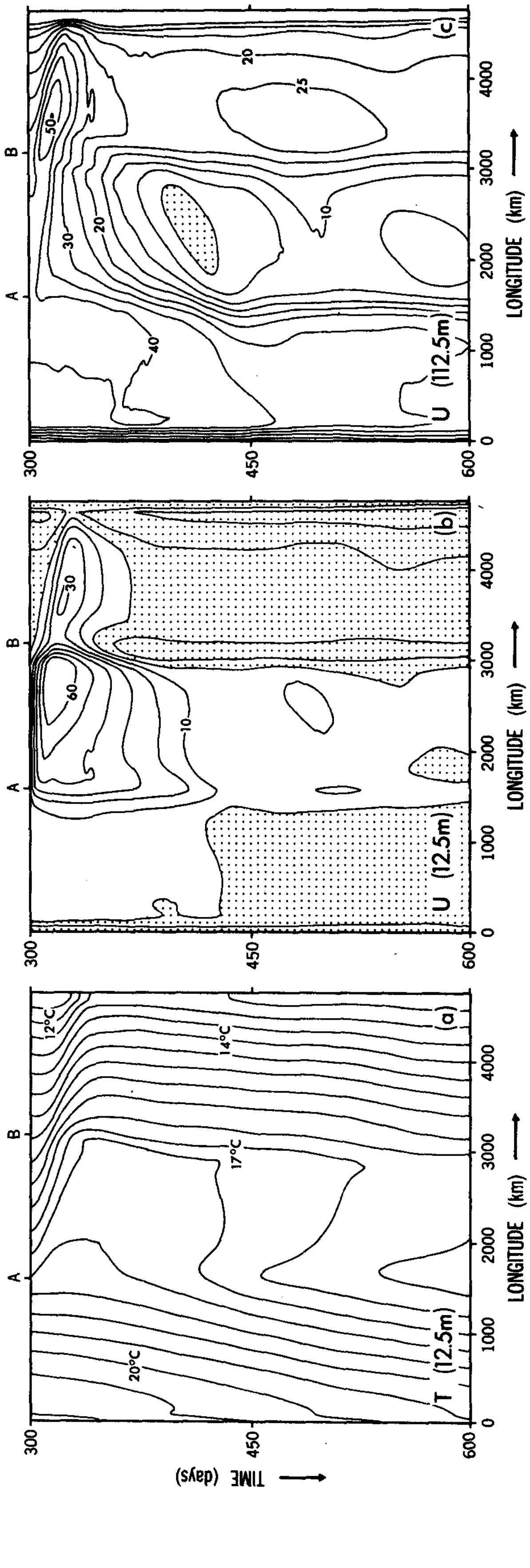

FIG. 6. As in Fig. 5, but the winds relax between meridians A and B only.

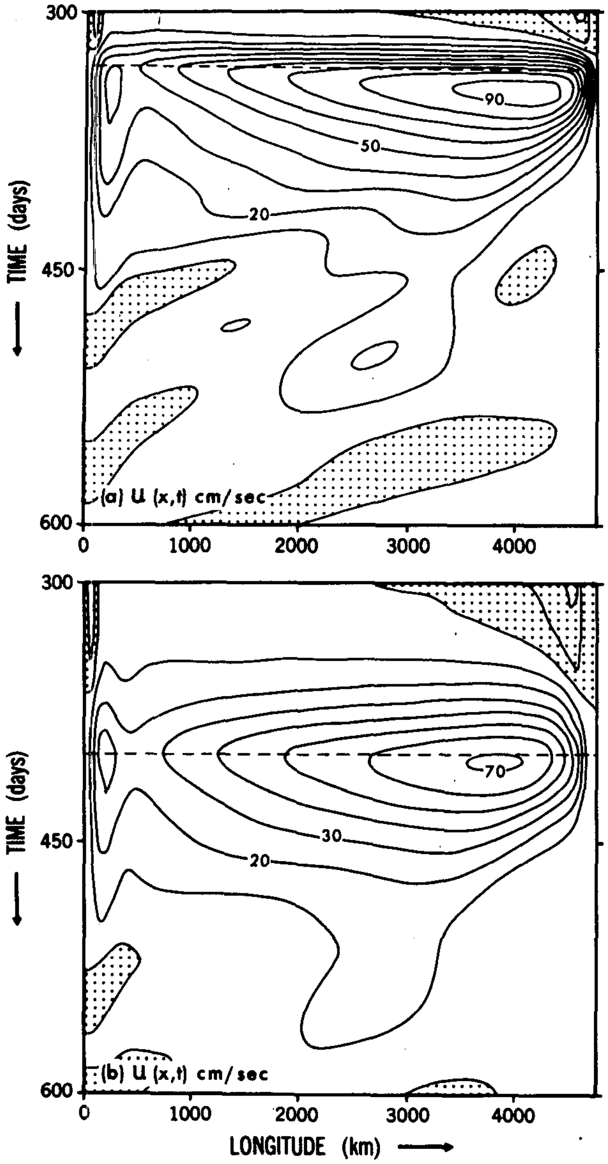

FIG. 7. The evolution of the surface flow along the equator when uniform westward winds relax gradually [according to Eq. (2)] over a period of (a) 30 days and (b) 100 days. The horizontal dashed line indicates when the wind stress is zero.

In Fig. 7a $T = 30$ days, in Fig. 7b $T = 100$ days. The longer the time over which the winds relax, the more difficult it is to discern the Kelvin waves. The acceleration of the surface currents does however remain pronounced. Hence, in practice, when the relaxation of the winds is erratic (in time and space) then identification of a Kelvin wave will be difficult but measurements should reveal an acceleration of the surface currents.

4. A temporary relaxation of the winds

A permanent relaxation of the winds results in a horizontal thermocline. This corresponds to a maximum transfer of heat from west to east. If the winds relax temporarily, for a time T, then there will be a stage when the thermocline is horizontal, provided T exceeds the adjustment time of the ocean. However, if T is less than the adjustment time of the basin then the thermocline may never be horizontal. Instead, a depression with a zonal extent that depends on T will propagate eastward along the equatorial thermocline. The amount of heat transferred eastward, and the warming in the east, will therefore be less than maximum. We expect the amplitude of the warming in the east to increase as T increases, and to asymptote towards a maximum value when T exceeds the adjustment time of the basin. This is shown schematically in Fig. 8.

When T exceeds the adjustment time of the basin then the oceanic response is practically in phase with the wind (see the discussion in Section 1). If however, T is short compared to the adjustment time then the maximum sea surface temperature in the east will occur a time T' after the winds stress had been zero. In other words, the nondimensional phase lag T'/T will decrease as T increases. This is shown schematically in Fig. 8.

To quantify the intuitive results described above we studied the oceanic response to winds that vary according to the expression

$$\tau^x = -0.5 + \alpha G(t)H(x), \tag{3}$$

where

$$G(t) = \exp[-4(t/T - 1)^2],$$

$$H(x) = 0.25\left[\tanh\left(\frac{x - 1600 \text{ km}}{120 \text{ km}}\right) - \tanh\left(\frac{x - 3200 \text{ km}}{120 \text{ km}}\right)\right],$$

$$\alpha = 1.$$

The winds relax only between meridians A and B, 1600 and 3200 km from the western coast respectively, for a time T approximately.

The results described in Section 3 suggest that the weakening of the winds will leave the region west of A unaffected. Between A and B the surface flow will accelerate eastward, but the core of the Equatorial Undercurrent will decelerate and the upper ocean will warm. East of B temperatures will increase, and surface and subsurface flow will accelerate.

Fig. 9, which shows changes along the equator when $T = 100$ days, confirms these inferences. (The vertical structure of the transients is shown in more detail in Fig. 10.) Note again that it is difficult to discern Kelvin waves. From Fig. 11 it is evident that for a different value of T—30 days in this case—the results are qualitatively similar. The amplitude of the warming, however, is only 2/3 as large as in the

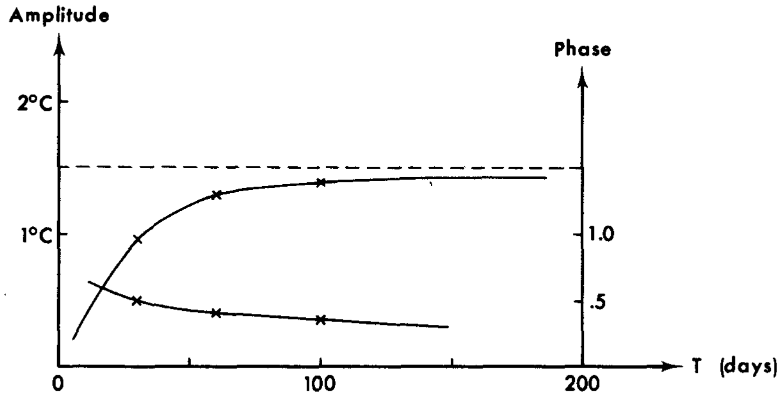

FIG. 8. A schematic diagram that shows how the warming at a point near the eastern coast increases as T, defined in Eq. (3), increases. The dashed line corresponds to an abrupt and permanent relaxation of the winds. The lower curve shows the (nondimensional) phase difference between the forcing and response.

case when $T = 100$ days and the non-dimensional phase lag T'/T is larger. The crosses on the curves in Fig. 8 show the points for which we have made calculations.

Fig. 8 gives the impression that the approach to an equilibrium response is very rapid as T increases and is of the order of 150 days for the basin being studied here. This is true if we are interested in sea surface temperature only. In an equilibrium response the currents too should vary in phase with the winds. A gradual relaxation of the winds, for example, should be accompanied by a gradual weakening of the currents. It is evident from Fig. 9 (see also Fig. 7) that on a time scale of 100 days the variations of the surface currents are not in accord with this description. As T increases, the acceleration of the surface jet decreases but in an equilibrium response there should be no acceleration at all. This only happens in time scales far greater than 150 days.

Finally, a comment on the role of Rossby waves in the adjustment of the equatorial oceans. In the introduction we explained how the Kelvin wave, and the gravest equatorially trapped Rossby wave, are the principal waves excited when the winds relax abruptly. As the time scale of the wind variations increases, the Kelvin wave continues to be important, but the gravest equatorially trapped Rossby wave becomes less and less important, and higher order Rossby modes become increasingly more important. These latter waves propagate slowly and are so seriously affected by mean currents that their role in the adjustment of the equatorial ocean is unimportant. For a detailed discussion of this matter see Philander and Pacanowski (1980b).

5. The appearance of eastward winds

The previous two sections concerned the relaxation of westward winds. Over the western Pacific Ocean, however, the westward winds not only die out, but they reverse direction and at times are eastward for prolonged periods. According to linear theory the oceanic response to eastward winds is merely the negative of the response to similar westward winds. The nonlinear response, however, is dramatically different (Philander, 1979b; Cane, 1979). The reason for this is the meridional circulation. Eastward winds cause the surface waters to converge on the equator where there is downwelling and a suppression of the thermocline (the exceptionally deep mixed layer in the western equatorial Pacific in Fig. 1 must in part be due to the periodic spells of eastward winds there). The convergence in the surface layers is balanced by divergence at depth (before zonal variations and a zonal pressure force become important). The fluid that converges on the equator gains momentum directly from the wind, but the divergent fluid at depth transports little zonal momentum away from the equator. There is consequently a nonlinear intensification of the eastward surface jet at the equator. Westward winds, in contrast to eastward winds, cause equatorial upwelling and divergent surface flow which removes westward momentum from the equator so that the wind-driven westward current is weakened.

To investigate whether there is anything exceptional associated with a weakening of westward winds which then become eastward, we studied the oceanic response to winds that vary according to Eq. (3) with $\alpha = 1.4$. The winds, which are briefly eastward and which attain a maximum eastward stress of 0.2 dyn cm^{-2} on day 400, are shown in the left-hand panel of Fig. 12.

The response is qualitatively similar to the one shown in Fig. 9, but is much more intense. Even though the amplitude of the wind variations is only 40% larger, the temperature increase is 70% larger in Fig. 12 than in Fig. 9. The nonlinear eastward

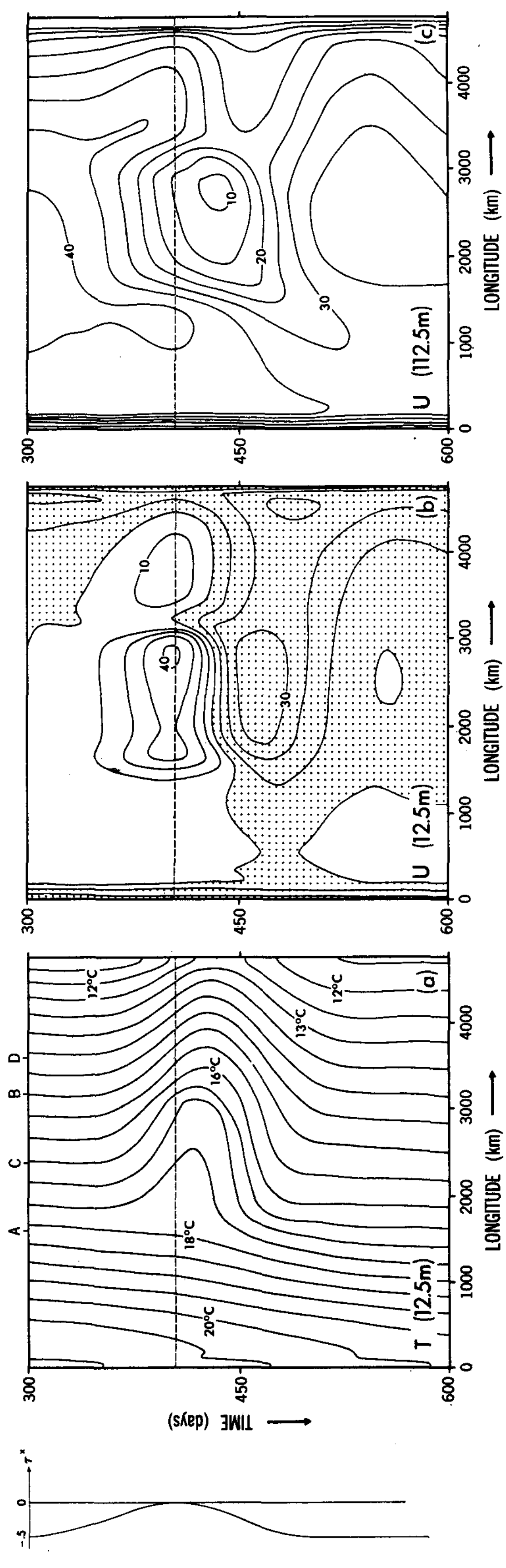

FIG. 9. As for Fig. 6 but the relaxation of the winds is temporary, for 100 days approximately (see the left-hand panel). The horizontal dashed line shows when the wind stress is zero.

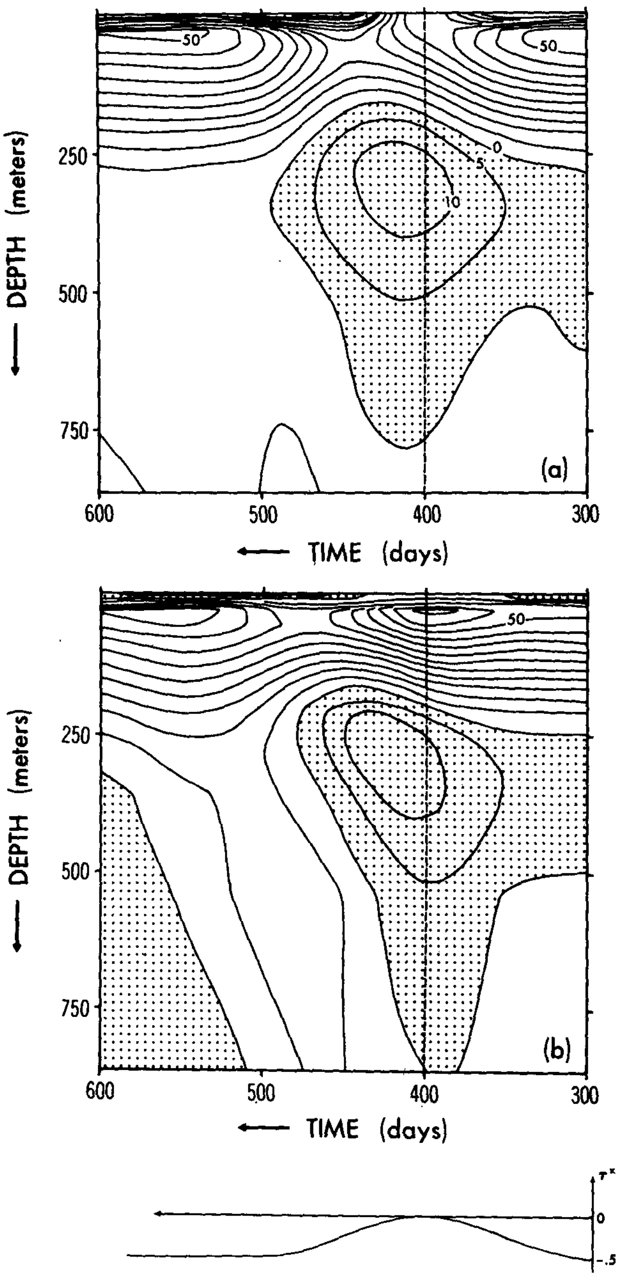

FIG. 10. Zonal current variations (cm s^{-1}) [westward flow shaded] as a function of depth and time, at (a) point C and (b) point D on the equator. Points C and D are indicated in Fig. 9. Wind variations, shown in the bottom panel, are the same as in Fig. 9. The wind stress is zero along the vertical dashed line.

intensification of the surface flow causes a more efficient eastward transport of heat. The eastward flow is furthermore convergent so that the jet draws heat from the surface layers of a latitudinally extensive region. The appearance of eastward winds over the western Pacific could therefore cause a very prolonged El Niño in the eastern equatorial Pacific Ocean.

6. Summary and discussion

A relaxation of westward winds has been shown to result in a zonal redistribution of heat, and consequently higher sea surface temperatures in the eastern part of the equatorial basin. The increase in temperature is proportional to the zonal extent of the region over which the winds relax and can be as large as 5°C if winds of intensity 0.5 dyne cm^{-2} die out completely over a 5000 km area. If the winds relax over a region 1600 km wide then the temperature increase is ~1.5°C. The increase in temperature also is a function of the duration T of the weakening of the winds. As T increases, the warming in the east increases until it asympotes to a maximum when T exceeds the adjustment time of the basin. (The maximum value is associated with a complete and permanent relaxation of the wind.) Fig. 8 shows how the amplitude of the warming, and the phase lag between forcing and response, depend on T. The scale of the abscissa in Fig. 8 is appropriate for sea surface temperature variations in a basin 5000 km wide. (The currents approach an equilibrium response much more gradually.) In the case of the much wider Pacific Ocean the time scale is approximately three times longer so that the winds would have to relax for ~400 days for maximum heating to occur there.

If the westward winds should die out and reverse direction, then the appearance of eastward winds for a brief period will cause a disproportionately large increase in sea surface temperatures in the eastern part of the basin. (This is associated with nonlinear effects discussed in Section 5.)

If winds relax locally then advection by equatorial currents is responsible for the zonal redistribution of heat. Should the time scale of the relaxation be less than the adjustment time of the basin then this advection is associated with an eastward acceleration of the surface waters. As the time scale of relaxation increases, the acceleration becomes less pronounced. At low frequency the means by which heat is redistributed is practically undetectable because it is due to a slight imbalance between eastward and westward currents.

Winds that relax over a part of the basin only leave the equatorial region to the west nearly unaffected. The equatorial region to the east is affected by Kelvin waves from the extremes of the forced region. They introduce a warming, and an eastward intensification of currents there. This warming is independent of longitude. (A local weakening of the winds by contrast, causes a temperature rise which increases in an eastward direction.) The Kelvin waves are discernible when the winds relax abruptly over a region with clearly defined boundaries. If the winds die out erratically, over a period of several months, and over a region with ill-defined boundaries, then Kelvin waves associated with the adjustment of the ocean

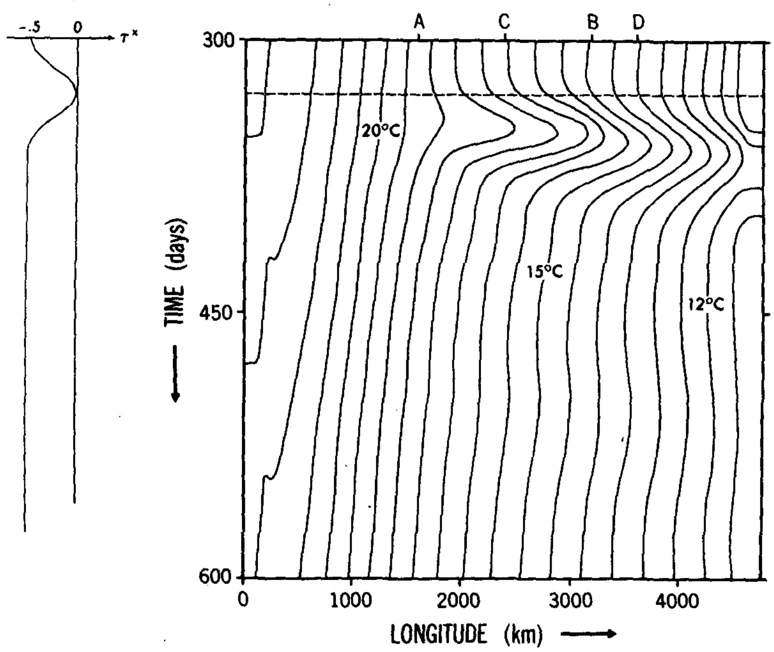

FIG. 11. Temperature variations along the equator when the winds weaken for ~30 days (=T). The horizontal line shows when the wind stress is zero.

to large-scale changes in the wind are not directly identifiable. (We are here concerned with Kelvin waves due to specific changes in the wind, so that the problem is a deterministic one. Of course, it should be possible to identify the Kelvin waves that form part of the spectrum of waves excited by stochastic fluctuations of the wind at periods of days and weeks.)

Analysis of the available data sets that describe the various El Niño events should provide a check on the results presented here. A comparison of the different events, for example, should reveal to what extent this phenomenon is affected by the length of time for which the winds relax, by the zonal extent of the region over which the winds relax, and by the appearance of eastward winds over the western

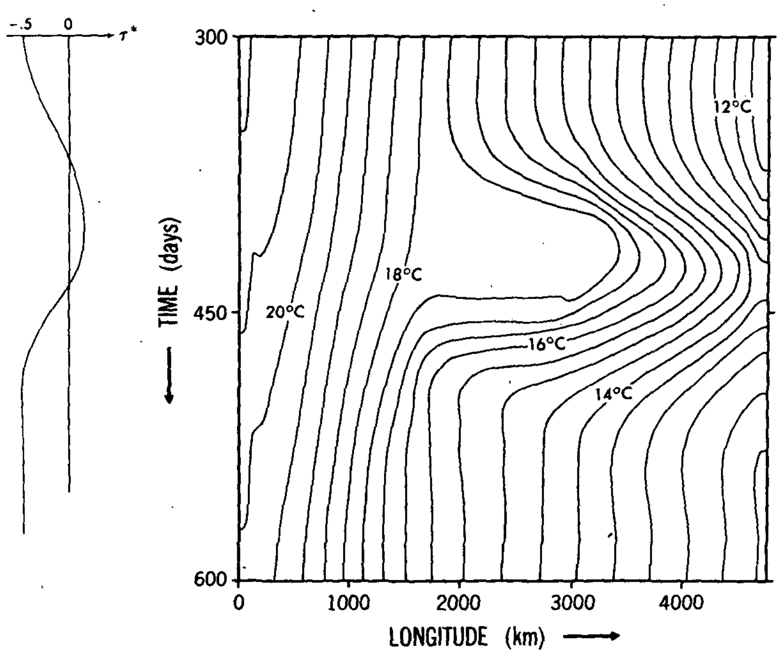

FIG. 12. As in Fig. 11 but the winds are now eastward for a brief period, as shown in the left-hand panel.

part of the Pacific. There clearly is a need for further measurements. Of particular value will be time series measurements of currents and subsurface temperatures to determine how the currents change when the warming occurs.

Most theoretical studies (this one, for example) concern only the later stages of El Niño, and concern only the region within a few degrees latitude of the equator. The anomalously warm surface waters which appear during El Niño are evident before the trades relax, and are not confined to the immediate vicinity of the equator. Wyrtki (1975) furthermore, has demonstrated that extra-equatorial currents such as the North Equatorial Countercurrent and South Equatorial Current change significantly during El Niño. The theoretical efforts therefore focus on only one (possibly small) aspect of El Niño, associated with changes in the intensity of the zonal wind stress close to the equator. Presumably variations in the intensity of the meridional winds (see Philander and Pacanowski, 1980c) and variations in the curl of the wind stress can also induce anomalous sea surface temperatures in the eastern equatorial Pacific Ocean. Further theoretical and observational studies of the variability of the tropical currents are necessary for an understanding of El Niño.

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