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The oceanic response to cross-equatorial winds (with application to coastal upwelling in low latitudes)

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ABSTRACT

Sea surface temperature variations observed in the eastern equatorial Atlantic and Pacific Oceans on seasonal, and possibly interannual (El Niño) time scales, may to a large extent be due to the variability of the local meridional winds. In a numerical model of the ocean, southerly winds cause low sea surface temperatures in the southeastern part of the basin because the coastal upwelling zone is extended far westward by (1) advection and (2) Rossby wave propagation which is important on time scales greater than a month. North of the equator sea surface temperatures are high. The thermocline has a trough near 3° N where there is an intense eastward jet. A relaxation of the southerly wind causes a warming in the southeastern part of the basin primarily because of a zonal redistribution of heat by the South Equatorial Current and Countercurrent.

1. Introduction

Winds parallel to a coast can cause intense coastal upwelling (Charney, 1955) which is confined to the neighbourhood of the coast only on time scales short compared to the period of Rossby waves. The propagation of these waves away from the coast, into the oceanic interior, widens the upwelling zone enormously and makes it a nonlocal phenomenon (Anderson and Gill, 1975). Since internal Rossby waves are possible at periods as short as a few months in the tropics, it follows that the seasonal upwelling off the coasts of Peru, north and southwest Africa, and northeast Africa cannot be simulated as local phenomena.

In addition to Rossby waves, advection by currents perpendicular to a coast will modify the coastal upwelling on long time scales. This is a particularly important factor in low latitudes because cross-equatorial winds (which induce upwelling along north-south coasts) are partially responsible for the eastward North Equatorial Countercurrent (north of the equator) and the westward South Equatorial Current observed south of the equator (Moore and Philander, 1977; Cane, 1979).

This paper is a numerical study of the oceanic currents and the temperature field, induced by variable cross-equatorial winds. The simulated sea surface temperature changes induced by gradually varying winds are remarkably realistic when compared to the seasonal variations observed in the eastern equatorial Atlantic and Pacific Ocean where the winds are predominantly from the south. This result suggests that a weakening of the meridional winds can be an important factor in the occurrence of some El Niño events in the eastern equatorial Pacific Ocean.

Section 2 of this paper describes the numerical model to be used here. Section 3 concerns the response to a sudden intensification of southerly winds, Section 4 the response to periodic winds. The results are summarized and discussed in Section 5.

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

$$\rho = \rho_0(1 - \alpha T)$$

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where ρ is the density, T is the temperature, $\alpha = 0.0002/°C$ is the coefficient of thermal expansion and $\rho_0 = 10^{-3}$ kg/m^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

$$\nu_v = 10^{-3}\,\mathrm{m^2/s};\quad \nu_H = 2\times 10^3\,\mathrm{m^2\,s};$$
$$\kappa_v = 10^{-4}\,\mathrm{m^2/s};\quad K_H = 10^3\,\mathrm{m^2/s}$$

Philander and Pacanowski (1980) 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 km and 70 km in the latitudinal and longitudinal directions respectively). In the vertical there are 16 grid points. The first 11 are spaced irregularly as shown in Fig. 1.

Motion is forced at the ocean surface ($z = 0$) by imposing a meridional windstress τ^y and a heat-flux (T_z). The zonal component of the windstress is zero.

$$\nu U_z = 0;\quad \nu V_z = \tau^y;\quad w = 0,$$
$$T_z = 10^{-4}\qquad (T - 25\,°\mathrm{C})$$

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. 1. 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 long compared to a decade (see Philander and Pacanowski, 1980). Here we concern ourselves with periodic 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. The imposed heat flux at the ocean surface reduces the diffusive cooling there and causes only a very slight warming of the deep ocean.

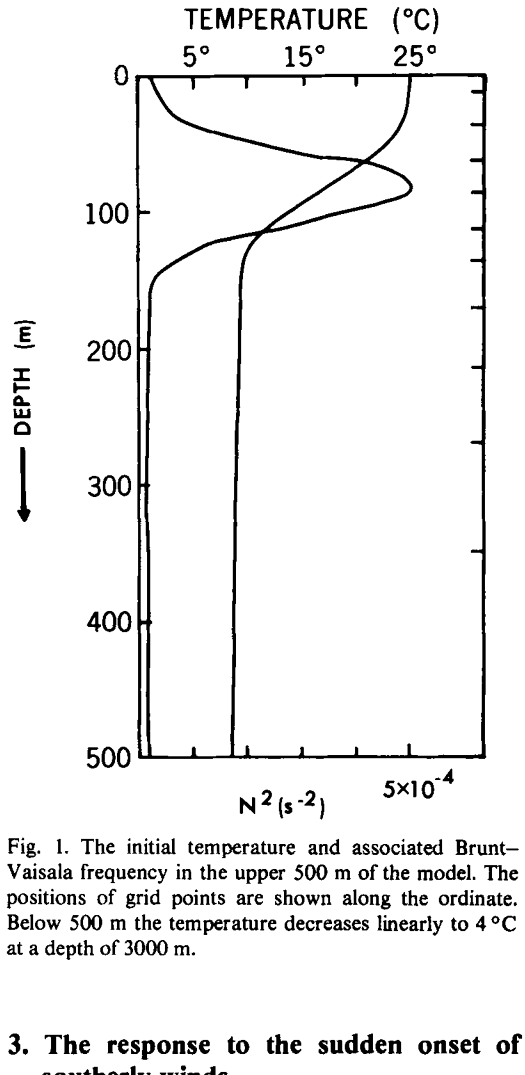

Fig. 1. The initial temperature and associated Brunt–Vaisala frequency in the upper 500 m of the model. The positions of grid points are shown along the ordinate. Below 500 m the temperature decreases linearly to 4 °C at a depth of 3000 m.

3. The response to the sudden onset of southerly winds

Moore (see Moore and Philander, 1977) first studied the generation of zonal currents by cross-equatorial winds in a linear shallow-water model. The important parameters are the following: the gravity wave speed $c = (gh)^{1/2}$; the equatorial radius of deformation $\lambda = (c/\beta)^{1/2}$; and the inertial period $T = 2\pi/(\beta c)^{1/2}$ at a distance λ from the equator. (Here g is the gravitational acceleration, h is the equivalent depth of the model and β is the latitudinal derivative of the Coriolis parameter f.) Moore showed that, within a time T of their onset, winds from the south generate an eastward current north of the equator, and an equal and opposite current south of the equator. The cores of these currents are at a distance λ from the equator. This is about 300 km for oceanographic values of the parameters. After a time T (approximately 1 week) equilibrium conditions are attained because a meridional pressure gradient exists to partially balance the wind stress. These results are valid in the absence of meridional coasts. In the presence of these coasts the response to constant southerly winds τ^y is a different equilibrium state in which the ocean is motionless and a meridional pressure gradient P_y balances the wind:

$$\frac{1}{\rho_0} h P_y = \tau^y \quad (1)$$

The zonal currents generated initially disappear completely! Cane and Sarachik (1979) describe the evolution of this new equilibrium state lucidly. Initially Rossby waves are excited at the north–south coasts. Those that propagate eastward are unimportant because their group velocities are so small that they remain in the vicinity of the western coast for a long time. The long nondispersive waves excited at the eastern coast propagate rapidly (with speed $c/5$, $c/9$, $c/13$ ...) and leave in their wake a state of no motion and the pressure gradient given by (1).

If the effect of nonlinearities are taken into account then the final equilibrium state can be completely different because the currents generated initially persist in modified form (Cane, 1979). In response to winds from the south the eastward jet north of the equator is intensified, the westward flow south of the equator is weakened. This result suggests that the observed North Equatorial Countercurrent may in part be due to the southerly component of the wind stress. It is intriguing that seasonal changes in the intensity of this current (Wyrtki, 1974) are in phase with seasonal changes in the intensity of the southerly winds at the equator: amplitudes are small in the spring when the ITCZ (Intertropical Convergence Zone) is close to the equator, and are large in the autumn when the ITCZ is in its northernmost position.

The various studies referred to above do not resolve the vertical structure of the motion because the investigators used either one- or two-level models. Such models also make it difficult to infer the sea surface temperature changes induced by variable winds. The multi-level model described in Section 2 simulates the continuous stratification of the ocean and thus overcomes these problems.

Fig. 2 shows the evolution of the zonal currents in our nonlinear model in response to the sudden onset of winds of intensity 0.05 N/m^2 from the south. The flow changes rapidly within the first week, but changes only gradually thereafter. The initial adjustment is associated with the establishment of a meridional pressure gradient to partially balance the wind. During this stage, the flow is independent of longitude so that the meridional circulation is closed: northward motion in the surface layers is compensated by an equal southward transport in the thermocline. The associated advection of momentum causes the westward motion of the surface layers in the southern hemisphere to penetrate into the northern hemisphere.

The structure of the oceanographic fields after the intitial adjustment is shown in Fig. 3. The eastward jet north of the equator is practically the only feature common to a linear and nonlinear model. This jet is associated with downwelling which causes the thermocline to be suppressed and pinched to the north of the equator. South of this region upwelling causes isotherms to slope upwards. The slope implies westward geostrophic flow north of the equator, but eastward geostrophic flow south of the equator. Hence the various currents in Fig. 3 appear to be in approximate geostrophic balance. The subsurface eastward current south of the equator is an intriguing feature. Calculations with a more realistic wind stress (which has a zonal component) are necessary to determine whether this is a simulation of the subsurface eastward "countercurrent" observed in the Pacific and Atlantic Oceans by Tsuchiya (1975) and Hisard, Citeau and Morliere (1976).

Over most of the ocean basin the flow is initially independent of longitude and has the characteristics described above. Along the northwestern and southeastern coasts, however, the winds induce upwelling. Rossby waves excited at the coasts, and advection by the zonal currents described earlier, extend the upwelling zones into the oceanic interior. The gradual changes that occur in Fig. 2 after the first week are associated with coastal effects that penetrate into the oceanic interior. The extent of these changes after 200 days is evident in Fig. 4. Note that the entire southeastern part of the basin now has low temperatures. By contrast there is

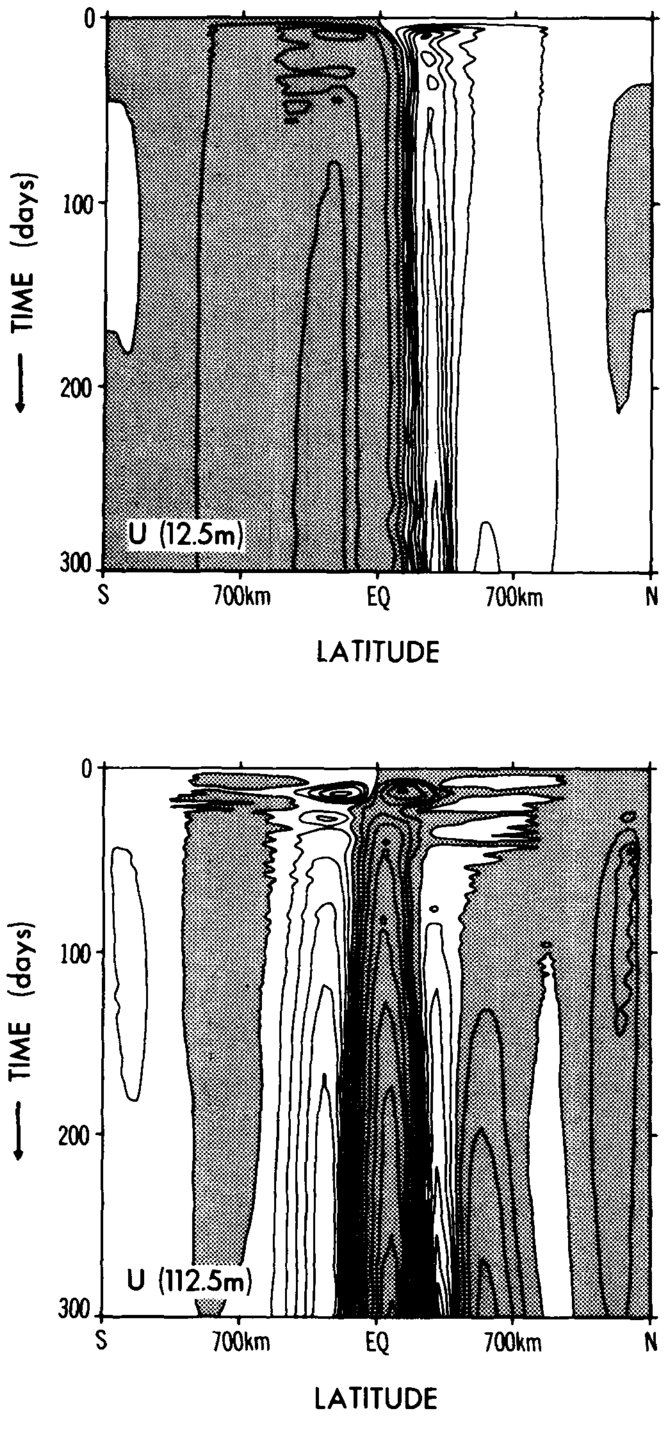

Fig. 2. The evolution of the zonal velocity component along the central meridian of the basin in response to winds from the south. The contour interval is 0.1 m/s at 12.5 m and 0.02 m/s at 112.5 m. Areas of westward flow are shaded.

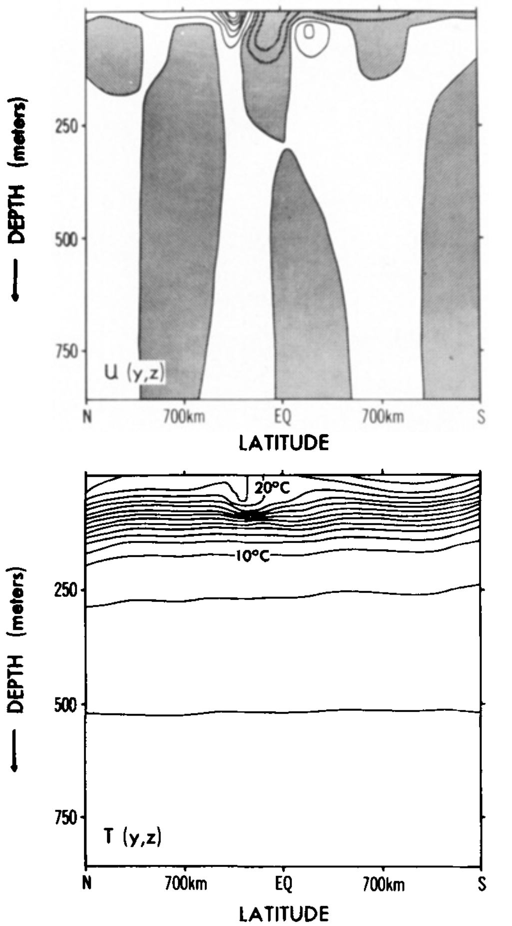

Fig. 3. A meridional section of the zonal velocity and temperature at day 200 along the central meridian of the basin. Contours are at intervals of 0.1 m/s; shaded areas designate westward flow.

only a relatively small tongue of cold water that extends eastward from the northwestern coast. (We remarked earlier that the long nondispersive Rossby waves excited at the eastern coast propagate into the oceanic interior much more efficiently than the short dispersive Rossby waves from the western coast.)

The cross-equatorial temperature differences in the eastern part of the basin are striking and suggest that the meridional component of the wind is responsible for similar differences in the eastern equatorial Pacific and Atlantic Oceans (see Fig. 5).

The stream function in Fig. 4 shows that an intense western boundary current, which is confined to the neighborhood of the equator, is fed by the western equatorial current, and in turn feeds the eastward surface current north of the equator. The cold "eddy" near 5° N is not unlike a similar feature observed off the east African coast. (For a more detailed study of the motion near the northwestern coast in a nonlinear shallow water model see Hurlburt and Thompson, 1976.)

The currents generated during the first 10 days are permanent features of the final equilibrium state. Rossby waves affect them to a far smaller degree in a nonlinear than in a linear problem. After the rapid adjustment during the first 10 days—this is essentially the time scale $2\pi(\beta c)^{-1/2}$ mentioned earlier—the currents accelerate gradually (see Fig. 2) until the flow becomes unstable. Loss of energy through instabilities then permits an equilibrium. The unstable waves have a wavelength of approximately 1000 km, a period of a month and cause the meandering evident in the isotherms near the equator in the eastern part of the basin in Fig. 4. The instabilities are associated primarily with the latitudinal shear between the eastward and westward surface currents in Fig. 3 and have been discussed by Philander (1978).

4. The response to fluctuating winds

The previous section concerned the attainment of equilibrium conditions after the sudden onset of southerly winds. Here we study the response to periodic winds. The southerly component of the observed winds have a particularly large amplitude over the eastern part of the equatorial Atlantic and Pacific Oceans. We therefore confine winds to the eastern part of the basin to determine to what extent those winds affect conditions in the west.

$$\tau^y = 0.05[1 + \sin(2\pi t/T)]H(x)\ \mathrm{N/m^2}$$

where

$$H(x) = 0.5 + 0.5 \tanh[(X - 2400\ \mathrm{km})/100\ \mathrm{km}]$$

Numerical integrations are continued for seven cycles of the periodic winds. Initially the ocean is motionless. We find that differences between the sixth and seventh cycles are very small compared to the differences that occur during a cycle. We therefore regard the seventh cycle as representative of a steady-state periodic solution.

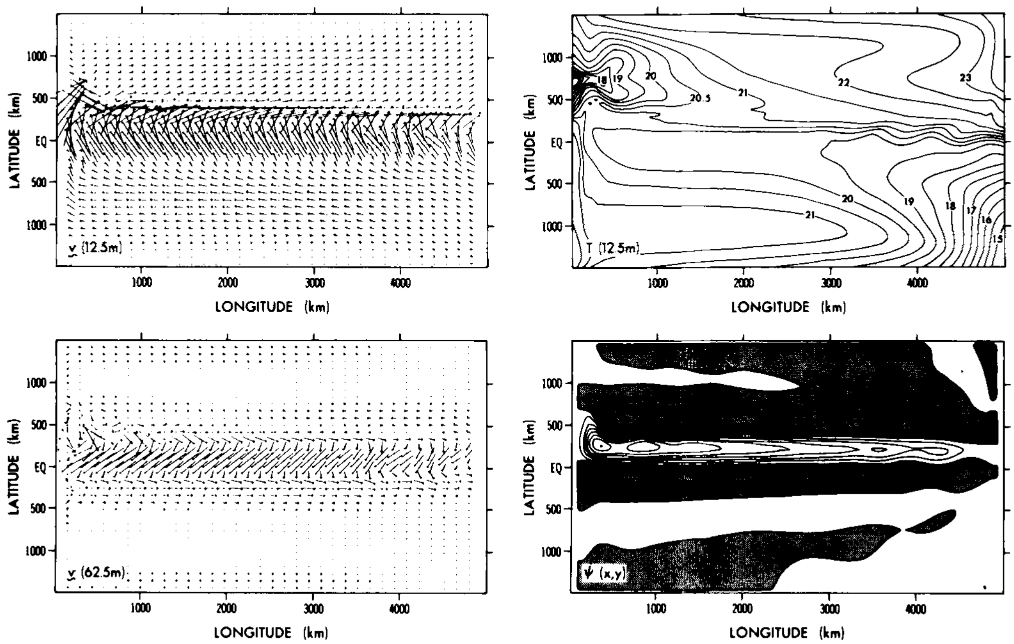

Fig. 4. The left-hand panels show horizontal velocity vectors at depths of 12.5 m (top) and 62.5 m (bottom). The right-hand panels show the temperature at a depth of 12.5 m and the vertically integrated transport stream function ψ. (Contours are in units of 10^6 m³/s.) These are instantaneous values at day 200.

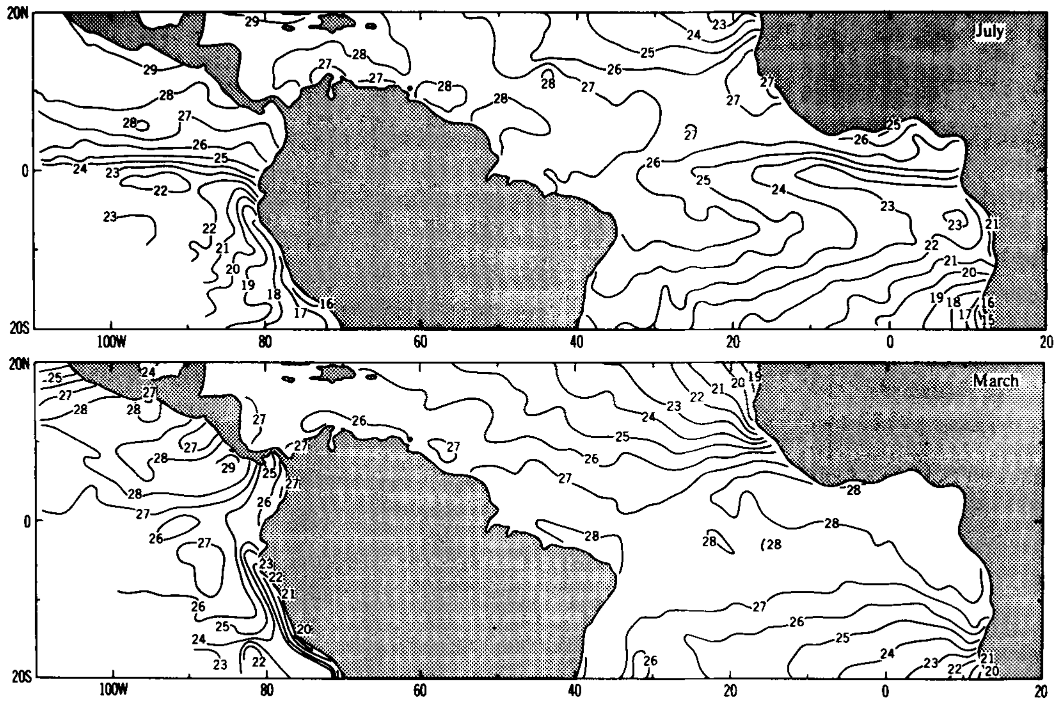

Fig. 5. Sea surface temperatures in the tropical oceans when the meridional component of the southeast trade winds is small (March) and when large (July). Isotherms are at intervals of 1 °C. After Hastenrath and Lamb (1977).

Since the currents in Fig. 3 are established within approximately 10 days, winds that oscillate with a period T smaller than 10 days will primarily excite (inertia-gravity) waves rather than currents. At periods longer than 10 days currents similar to those in Fig. 3 are of central importance in the oceanic response. The extent to which this response is influenced by Rossby waves excited at the eastern coast depends on the period T. No (antisymmetric) Rossby waves are possible at periods much shorter than 50 days, at longer periods these waves become increasingly more important. The Rossby waves affect the adjustment of the ocean to a change in wind conditions. Hence the longer the period T the closer is the ocean to an adjusted state. If the wind changes very gradually—on a time scale long compared to the adjustment time of the ocean—then the ocean is always in an adjusted state and is at each moment in equilibrium with the wind at that moment. On these long time scales the ocean has no "memory" of earlier winds. If the winds vary rapidly—on a time scale short compared to the adjustment time of the ocean—then the oceanic response is likely to be composed of two parts: a steady response induced by the mean winds plus a fluctuating component due to the oscillating winds. At these high frequencies the ocean "remembers" the mean value of the wind. These considerations explain why, in our numerical experiments, the amplitude of oceanic variability increases as the period T of the wind increases from 50 to 200 days. We find that the upwelling zone off the southeastern coast, as measured by the extent of water at a temperature of 18°C or less, varies relatively little at a period T of 50 days but at a period of 200 days (Fig. 6) changes are considerable.

A low frequency equilibrium response should be in phase with the forcing (because the ocean has no memory). It is evident from Fig. 6 that the response at a period of 200 days is not yet an equilibrium response. Although the currents vary practically in phase with the winds, changes in the density field lag. The area of the region in which the temperature is 18°C (or less) continues to grow after the winds reach their peak (at day 50) and continues to diminish after the winds are zero at day 150 (see Fig. 6). It is evident from the westward tilts in Fig. 7 that Rossby wave propagation is important in these fluctuations. The other important factor is advection. It is noteworthy that the westward flow south of the equator weakens when the winds relax, and that the eastward current south of the equator in Fig. 3 surfaces when the winds are very weak.

A comparison of the response at periods of 150 and 200 days reveals only small differences. The approach to the perfect equilibrium response mentioned earlier is apparently a very gradual one. Instabilities do not occur in our calculations because large shears are maintained for an insufficient length of time. (Our calculations were for periods less than 200 days. At longer periods instabilities are likely to be important, but only during the intense phase of the wind.)

Although the winds blow over the eastern half of the basin, significant motion, but only small temperature changes, are induced in the western unforced region (Figs. 6 and 7). The currents, which are due to westward propagating Rossby waves, are westward near the equator, and are eastward near 4° N (Fig. 6). The curl of the wind stress is such that, according to linear Sverdrup theory, there should be southward transport in a narrow meridional strip in the centre of the basin, northward transport along the western coast. It is evident that nonlinearities alter the stream function pattern significantly. Coastal currents such as those along the eastern coasts of Brazil and Africa, including the Somali Current, could be affected by the non-local meridional wind-stress component.

5. Summary and discussion

After the sudden onset of southerly winds over a motionless ocean, zonal currents with the structure shown in Fig. 3 are generated within 10 days. The eastward surface jet centred on 4° N is the most intense current. Motion at and south of the equator is westward but the subsurface flow near 3° S is eastward. The associated density field is consistent with a geostrophic balance: the thermocline is suppressed near 4° N (where the flow is convergent) and rises towards the surface in northward and southward directions. Upwelling is intense along the southeastern coast of the basin. Rossby wave propagation, and advection, extends the region of cold water far westward (Fig. 4). In the northwestern part of the basin a tongue of cold water (advected by the eastward current) extends eastward from the coast along approximately 5° N. A similar feature is associated with the Somali Current in the Indian Ocean (Bruce, 1980).

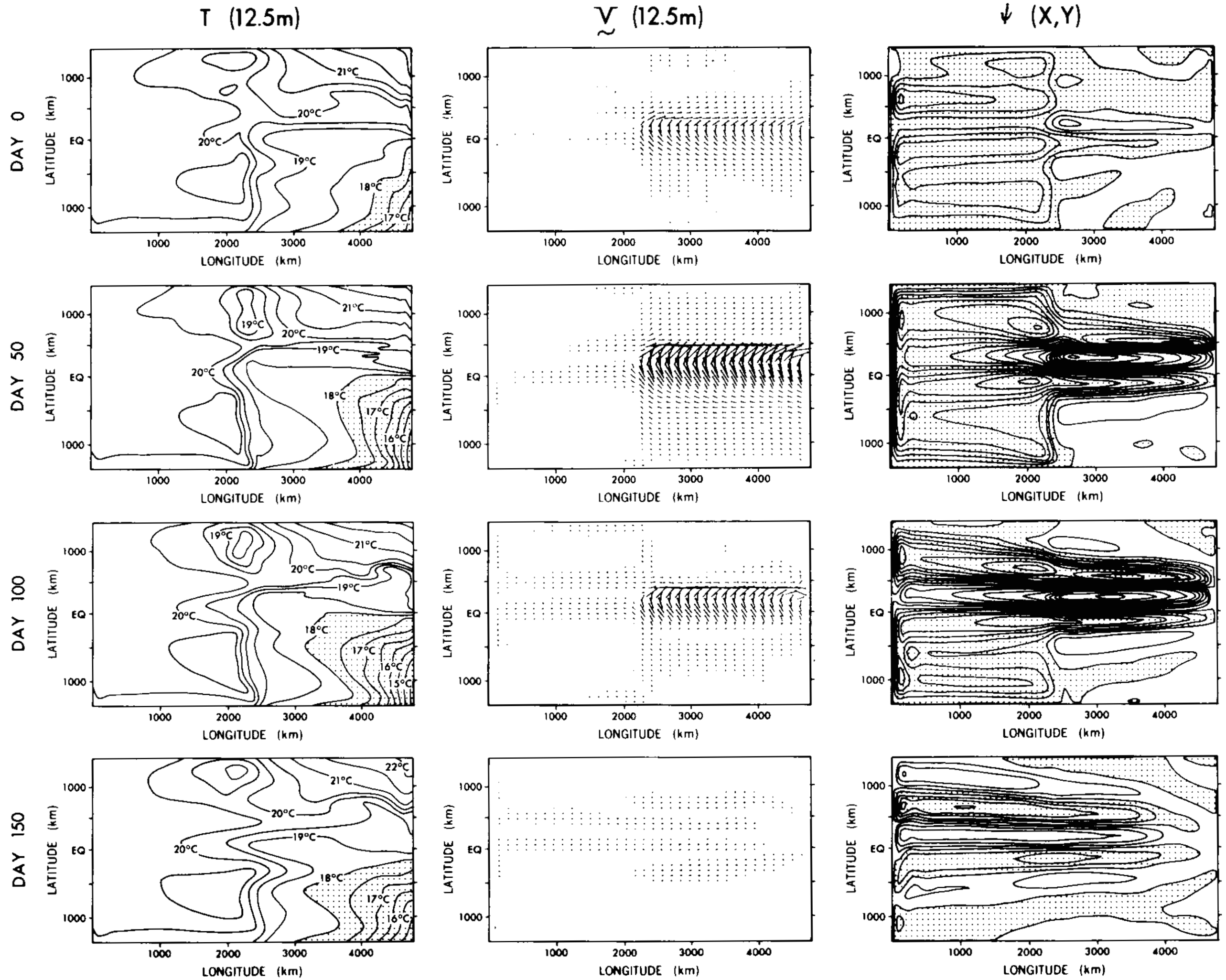

Fig. 6. Maps of the temperature T, horizontal velocity vector at a depth of 12.5 m, and vertically integrated horizontal transport steam function when the wind has a period of 200 days. The wind stress is 0.1 N/m^2 at day 50, zero at day 150.

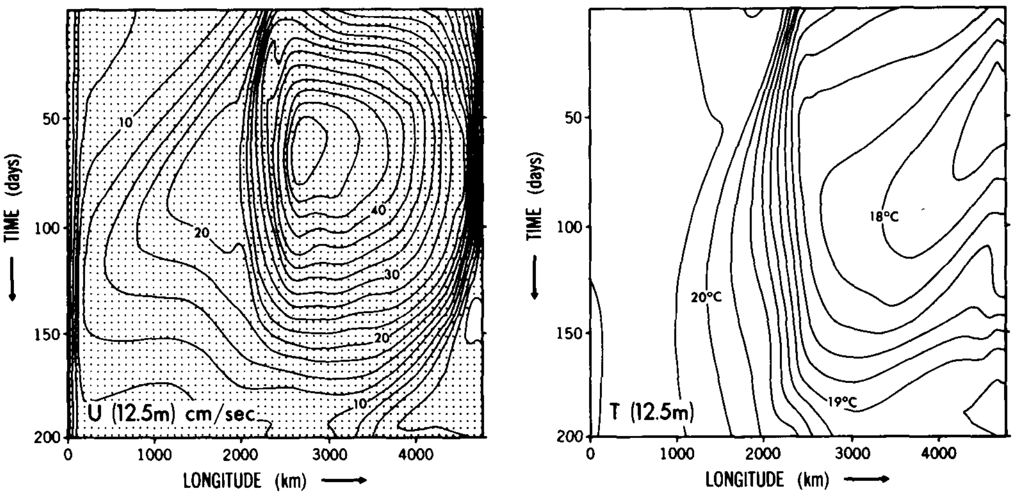

Fig. 7. Temperature and zonal velocity fluctuations along the equator when the periodic northward winds are confined to the eastern part of the basin.

If the winds are fluctuating then the amplitude of the variability increases as the period of the winds increases. At periods longer than 10 days the structure of the currents is essentially that shown in Fig. 3. The currents and wind vary almost in phase, but the density field lags: warming continues for a while after the wind has passed its weakest phase (Fig. 6). When the winds intensify, the upwelling zone in the southeastern part of the basin expands because of advection and Rossby wave propagation, as explained earlier. When the winds relax, the reverse happens. (The westward current south of the equator weakens, and the eastward current there surfaces when the winds weaken.) Consequently during one cycle there is a zonal redistribution of heat in the southeastern part of the basin. This is the primary cause of the sea surface temperature variations observed there, a cross-equatorial redistribution of heat is of less importance.

On the basis of the results described above we propose that variations in the intensity of the meridional winds (which prevail over the eastern equatorial Atlantic and Pacific Oceans) contribute significantly to the variability of sea surface temperatures in those regions, and are primarily responsible for the observed seasonal changes in sea surface temperature (compare Figs. 4 and 5). Over the Gulf of Guinea for example the southerly winds are weak in March, intensify from April until July and weaken gradually thereafter. According to our results sea surface temperatures south of the equator will be at a minimum a month or two after the winds are most intense. This is consistent with the data analysis of Merle et al. (1979) who, furthermore, show that the annual cycle of the sea surface temperature is associated with northwestward phase propagation.

If the meridional winds are seasonally important, then they are likely to be important on interannual time scales too. El Niño events, during which sea surface temperatures in the eastern equatorial Pacific Ocean are anomalously high, are indeed correlated with a weakening of the meridional winds (Hickey, 1975). This, however is only one factor that contributes to El Niño. A weakening of the zonal winds can be at least as important (Wyrtki, 1975; Barnett, 1977).

The available data are inadequate for a stringent test of the hypothesis that variations in the meridional winds significantly affect sea surface temperatures in the eastern equatorial Atlantic and Pacific Oceans. Of particular value will be an observational study of the seasonal cycle to determine how changes in the meridional winds, changes in the westward South Equatorial Current and eastward South Equatorial Countercurrent, and changes in the heat content are correlated. Further theoretical studies are necessary to take into account (i) more realistic spatial variations of the meridional winds, and (ii) the non-zero zonal component of the wind stress.

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ОТКЛИК ОКЕАНА НА ВЕТРЫ, ДУЮЩИЕ ЧЕРЕЗ ЭКВАТОР

Изменения температуры поверхности моря, наблюдаемые в восточных экваториальных частях Атлантики и Тихого океана с сезонными и, возможно, межгодичными (Эль-Ниньо) масштабами времени, могут быть в большой степени из-за изменчивости локальных меридиональных ветров. В численной модели океана южные ветры вызывают низкие температуры поверхности моря в юго-восточных частях бассейна, потому что зона берегового апвеллинга простирается далеко на запад под действием (1) адвекции и (2) распространения волн Россби, что важно на масштабах времени, больших месяца. К северу от экватора температуры поверхности моря высоки. Термоклин имеет ложбину вблизи 3°с.ш., где находится интенсивная западная струя. Ослабление южных ветров вызывает нагревание в юго-восточной части бассейна в основном из-за зонального перераспределения тепла Южным экваториальным течением и противотечением.