On the Contrast between the Seasonal Cycles of the Equatorial Atlantic and Pacific Oceans

S. G. H. PHILANDER AND YI CHAO

Atmospheric and Oceanic Sciences Program, Princeton University, Princeton, New Jersey

(Manuscript received 24 August 1990, in final form 19 February 1991)

ABSTRACT

Although the winds on the equator at 28°W in the Atlantic and 140°W in the Pacific have similar seasonal variations, the current fluctuations have pronounced differences. In the Pacific the maximum speed of the Equatorial Undercurrent, attained in the northern spring, can exceed 140 cm s^{-1}, while the minimum speed, in the autumn, is less than 80 cm s^{-1}. In the Atlantic the maximum speed of 80 cm s^{-1} hardly varies seasonally, although it tends to be largest in the autumn. Analyses of results from a realistic simulation of the equatorial currents indicate that the larger zonal extent of the Pacific, and the seasonal variations of the winds over the western Pacific, which can be out of phase with those in the east, are the principal reasons for the differences between the Atlantic and Pacific.

1. Introduction

Easterly trade winds prevail over most of the equatorial Atlantic and Pacific oceans. Seasonally, the winds at the equator are intense during the northern summer and autumn when the intertropical convergence zone (ITCZ) is farthest north and are weak in February, March, and April when the ITCZ is close to the equator. Although these seasonal variations are similar over the equatorial Atlantic and Pacific oceans, the responses of the two oceans to these winds have striking differences, which were recently discussed by Halpern and Weisberg (1989). Their measurements in Fig. 1 show that the maximum speed of the Equatorial Undercurrent hardly varies at 28°W on the equator in the Atlantic but is subject to a huge annual cycle at 140°W in the Pacific. (The two sites, 28°W and 140°W, are at approximately the same distance from their respective eastern coasts.) Another difference between the two oceans is in the vertical excursions of the thermocline as measured by the 20°C isotherm: it is much more pronounced at 28°W in the equatorial Atlantic than at 140°W in the Pacific. Halpern and Weisberg pose these differences as an enigma. In this paper we propose a solution.

The most obvious difference between the equatorial Atlantic and Pacific is in their longitudinal extents. This means that the time T it takes the oceans to adjust to a change in the winds is longer in the case of the Pacific than the Atlantic because T depends on the time it takes planetary waves to propagate across the ocean (Philander 1990). The width of the equatorial Atlantic is so small that its adjustment time is less than the seasonal time scale. This means that the response to the seasonal forcing should be an equilibrium one and should correspond to a succession of steady states. This result is confirmed by the measurements of Katz et al. (1977). They found that seasonal variations in the intensity of the zonal wind stress along the equator, and in the zonal pressure gradient maintained by the wind, are practically in phase in the western equatorial Atlantic.

The Pacific is far wider than the Atlantic and its adjustment time at the equator exceeds the seasonal time scale so that the response to seasonal forcing, unlike that of the equatorial Atlantic, should not be an equilibrium response. Philander (1979) suggested that whereas the zonal wind stress and pressure gradient are always in balance in the equatorial Atlantic, this should not be so in the equatorial Pacific. He proposed that when the winds along the equator relax during the northern spring, the pressure gradient fails to adjust rapidly in the Pacific, is no longer balanced by the wind, and accelerates the currents eastward, thus causing the high speeds during the spring at 140°W in Fig. 1. This argument is contradicted by the measurements of McPhaden and Taft (1988) who find an equilibrium response in the central and eastern equatorial Pacific; at the annual cycle, the zonal wind stress and the pressure gradient are essentially in balance. In other words, both the equatorial Atlantic and Pacific have an equilibrium response to the seasonally varying winds.

The flaw in Philander's (1979) arguments is the assumption that the width of the basin is the only length

Corresponding author address: Dr. George Philander, Program in Atmospheric and Oceanic Studies, Princeton University, P.O. Box CN 710, Sayre Hall, Princeton, NJ 08544-0710.

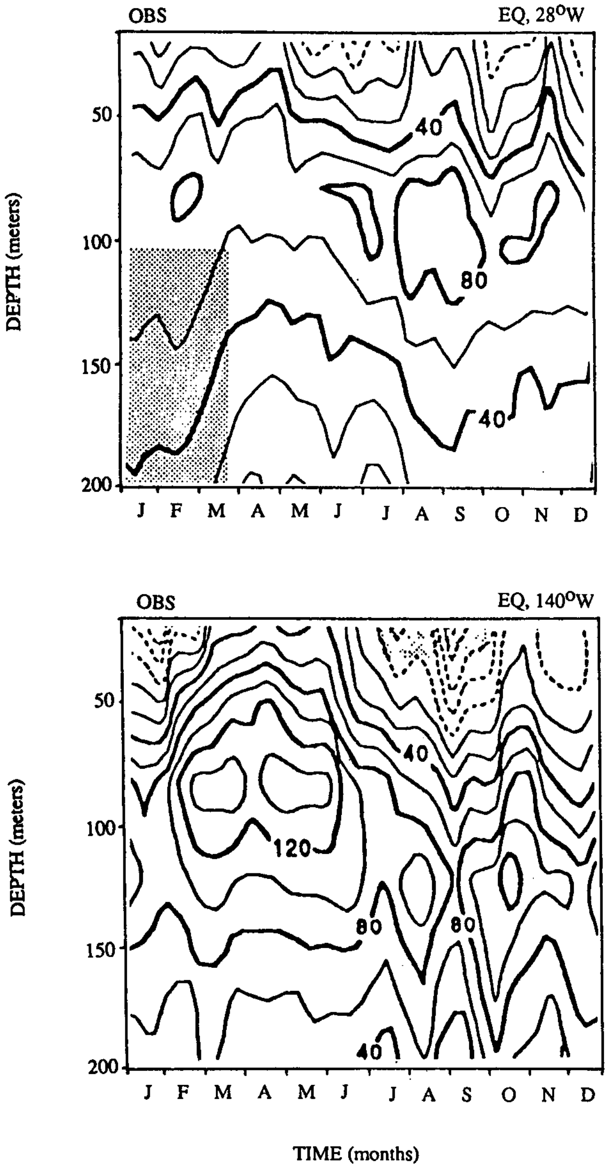

FIG. 1. Time series of the zonal currents (cm s^{-1}, westward where contours are dashed) as a function of depth (in meters) on the equator at 140°W in the Pacific and 28°W in the Atlantic Ocean. (After Halpern and Weisberg 1989.)

scale that determines the adjustment time. He overlooked another important length scale, the one imposed by the wind. The structure of zonal winds over the equatorial Atlantic and Pacific oceans have a very well-defined length scale because the annual cycle of these winds corresponds to a westward-propagating wave with a wavelength of approximately 15 000 km (Meyers 1979; Lukas and Firing 1985). This can be discerned in Fig. 2, which shows, not the annual harmonic, but climatological monthly mean values. In this figure the zonal wind fluctuations have a closely associated westward-propagating signal in sea surface temperature [first identified by Horel (1982)], but not in the meridional winds. Apparently there is more to the seasonal cycle of the surface winds near the equator than the north–south migrations of the intertropical convergence zone. It appears that interactions between the ocean and atmosphere are involved because seasonal changes in the sea surface temperature field, which influences the atmosphere, and in the zonal wind stress, which influences the ocean, are very similar. This topic will be discussed on another occasion. Here we focus on the response of the ocean to the seasonally varying zonal winds that have an east–west scale substantially smaller than the width of the Pacific.

To determine why the equatorial Atlantic and Pacific oceans respond so differently to the seasonally varying winds, we adopt the method of Wacongne (1989, 1990) and analyze the results from a numerical model that simulates the variability of the tropical oceans realistically. The next section briefly describes the model and the subsequent section describes the results.

2. The model

The oceanic model used in this study is the general circulation model developed by Bryan (1969) and Cox (1984) and modified for the tropical Pacific as in the study of Philander et al. (1987). The model has 27 levels in the vertical with 9 of those levels in the upper 300 m. The model ocean has realistic topography except for islands that are sunk 65 m below the surface. The grid spacing is 1.5° longitude and 1° latitude except ⅓° latitude between 10°N and 10°S. Vertical mixing is Richardson number dependent as in Pacanowski and Philander (1981). Horizontal eddy viscosity and diffusivity have the same value 1×10^7 cm^2 s^{-1}. The initial density field corresponds to the data of Levitus (1982). The model is first spun up, with the climatological monthly mean winds from COADS, for a 3-year period. Thereafter it is forced with the monthly mean surface winds from COADS as prepared by Oort et al. (1987) for the period 1967 to 1979. The surface heat flux is based on bulk formulas that use the model sea surface temperature and specified air temperature, humidity, and wind speeds. Data from the last 13 years of the simulation were used to calculate a climatological seasonal cycle for analysis. See Chao (1990) for details.

The model results agree with the few scattered measurements (primarily of sea level and sea surface temperature) that are available for the period 1967–79. Compare, for example, Fig. 1 with Fig. 3, which shows the climatological zonal current fluctuations on the equator at 140°W in the model. The Atlantic version of the model analyzed by Wacongne (1989) is similarly realistic so that we proceed with an analysis similar to that of Wacongne.

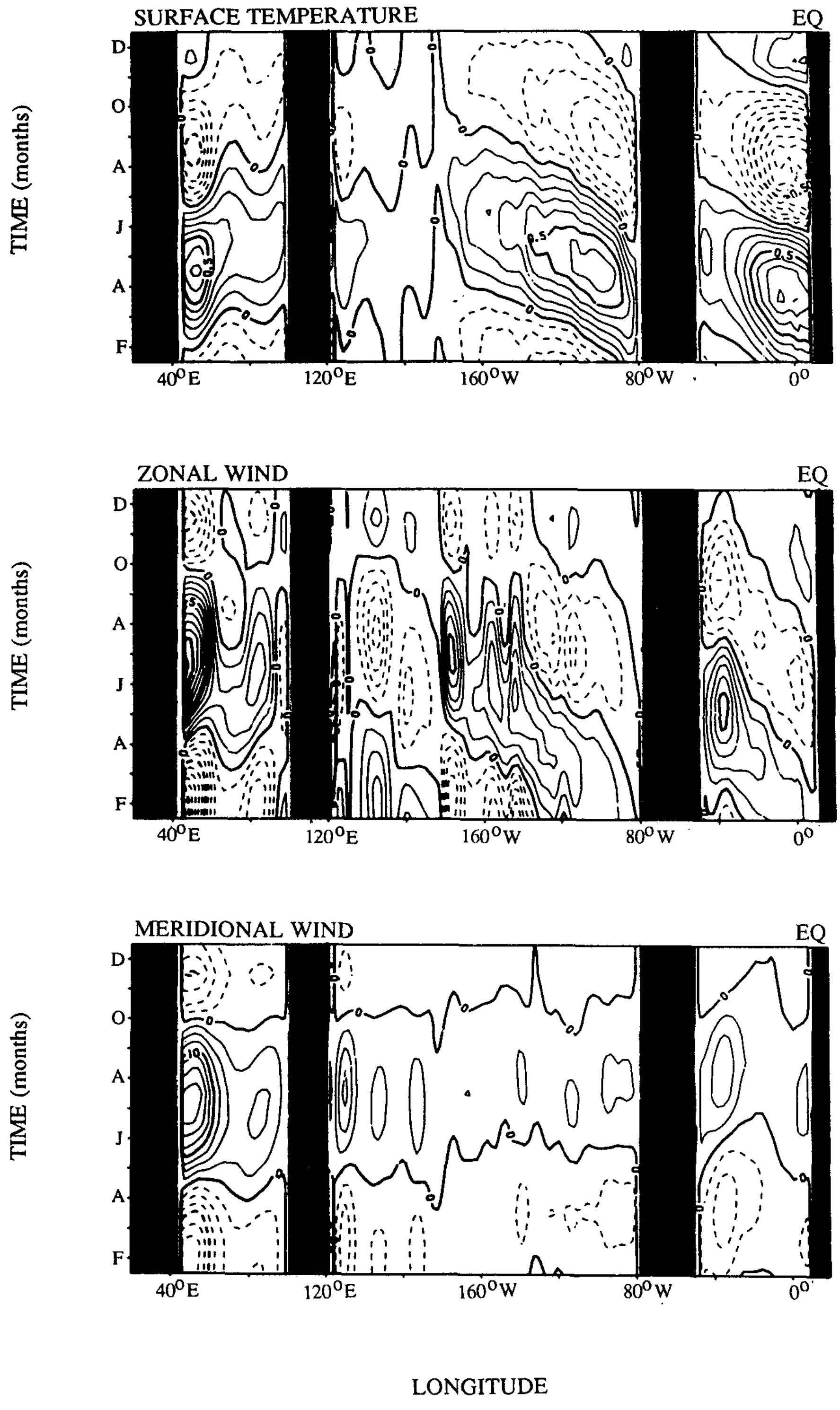

FIG. 2. Climatological monthly mean anomalies, along the equator, of sea surface temperature (contour interval: 0.25°C), zonal component of the surface wind (contour interval: 0.01 dyn cm^{-2}) and meridional component of the surface wind (contour interval: 0.025 dyn cm^{-2}). Dashed contours indicate negative, easterly, and northerly anomalies, respectively.

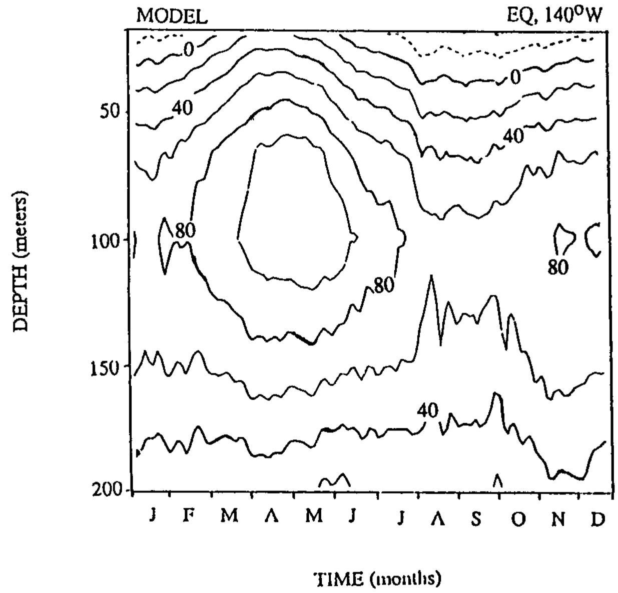

FIG. 3. Simulated zonal currents on the equator at 140°W. The contour interval is 20 cm s^{-1} and dashed lines indicate westward currents. Compare with the bottom panel of Fig. 1.

3. Results

Wacongne's (1989) analysis of the zonal momentum balance along the equator in the Atlantic identifies the regimes shown in Fig. 4 for annual mean conditions. The upper ocean, region I, is the most complex because there nonlinear acceleration is influenced strongly by both the zonal pressure gradient p_x and by the downward diffusion of momentum gained from the wind.

I:

$$VU_y + WU_z \approx -\frac{1}{\rho_0} p_x + (\nu U_z)_z. \qquad (1)$$

Here (U, V, W) are the velocity components in the eastward x, northward y, and upward z directions. This equation is for the equator where $y = 0$. Vertical mixing is denoted by ν. The notation is standard. The dynamical balance is simpler in region II in the west where the thermocline is deep and where the pressure force directly accelerates the Equatorial Undercurrent.

II:

$$UU_x + VU_y + WU_z \approx -\frac{1}{\rho_0} p_x. \qquad (2)$$

If a vertical integral is taken over this region, the term UU_x dominates the left-hand side. Immediately to the east of region II the thermocline shoals more rapidly than the core of the undercurrent does so that the undercurrent in region III is below the region where the eastward pressure force is large. The undercurrent in region III is therefore an inertial jet decelerated by dissipation.

III:

$$UU_x \approx A\nabla^2 U. \qquad (3)$$

Here A denotes horizontal eddy viscosity.

The zonal pressure gradient accelerates the Equatorial Undercurrent in region II so that it attains its maximum speed near the eastern extreme of this region. It follows that not only the magnitude of the pressure gradient but also its fetch determines the

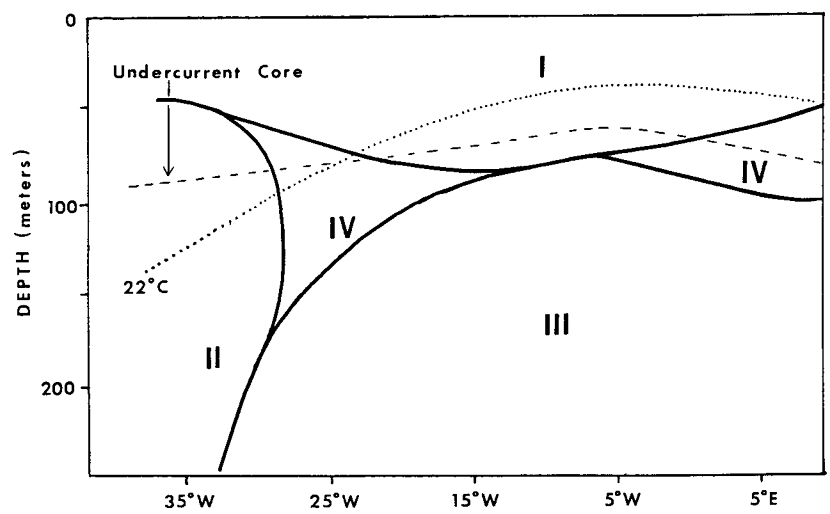

FIG. 4. Regions of different balances for the zonal momentum equation in the equatorial plane of the Atlantic Ocean. See Eqs. (1)–(3). (After Wacongne 1989.)

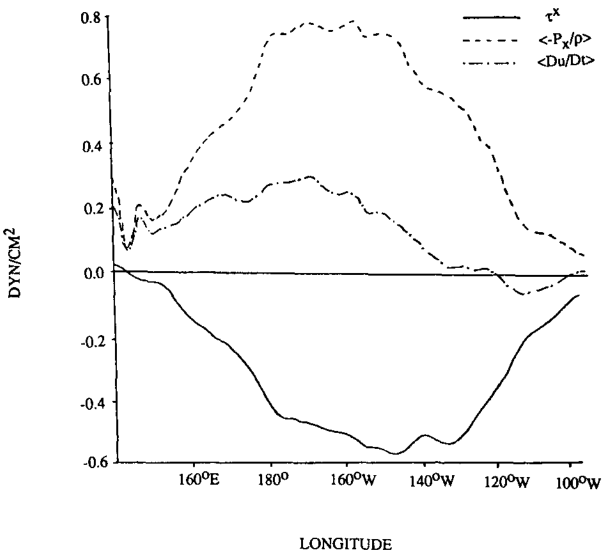

FIG. 5. The annual mean values, as a function of longitude along the equator, of the zonal wind stress τ^x, of the zonal pressure gradient integrated vertically over the upper 300 m of the model, and of the absolute current acceleration also integrated vertically.

maximum speed of the undercurrent. If region II covered a wider range of longitudes, the undercurrent would be more intense. Wacongne (1990) argues persuasively that this is the case in the Pacific Ocean. The crucial factors that determine why the maximum speed of the time-mean Equatorial Undercurrent is greater in the Pacific than the Atlantic are 1) the greater width of the equatorial Pacific and 2) the structure of the zonal winds in the Pacific, which maintain a zonal pressure gradient across the width of the basin. In the absence of mean easterly winds over the western half of the equatorial Pacific, the undercurrent there would be considerably weaker. Figure 5 confirms that the nonlinear acceleration terms are most important over the western half of the equatorial Pacific Ocean. This is in contrast to the region east of 140°W where the westward wind stress is essentially balanced by the vertical integral of the eastward pressure force while nonlinearities are of secondary importance.

The two factors that explain why the Equatorial Undercurrent is more intense in the Pacific than the Atlantic also explain why this current has different seasonal variations in the two oceans. The seasonal fluctuations of the zonal component of the wind are similar in the two oceans. However, since the annual cycle of the wind corresponds to a westward-propagating wave with a wavelength of approximately 15 000 km, the equatorial Pacific can accommodate an entire wave but the equatorial Atlantic cannot. This means that, in the equatorial Pacific, there are times when easterly wind anomalies prevail over one part of the basin while westerly anomalies prevail over another part. During the late northern summer, for example, the easterlies are particularly intense in the east, but westerly anomalies are present to the west. During the northern spring the situation is reversed. In Fig. 6 it is evident that at these extremes of the seasonal cycle the anomalous zonal pressure gradient can have opposite signs in different parts of the equatorial Pacific. Figure 7 depicts the change, between April and October, in the extent and location of region II in which this pressure gradient accelerates the Equatorial Undercurrent as indicated in Eq. (2). The contour inside region II shows where the acceleration is most intense. From this figure it follows that at 140°W the maximum speed of the Equatorial Undercurrent is greater in April than October because the magnitude and the zonal extent of the pressure force that accelerates the current is greater in April. In other words, to explain the current variations at 140°W it is necessary to take into account not only the winds at that location but also the winds farther west.

It is even easier to explain variations in the speed of the Equatorial Undercurrent in the Atlantic Ocean because the region of acceleration, region II of Fig. 4, has a small longitudinal extent. Hence the maximum speed of the undercurrent should coincide with the period of intense easterly winds near 28°W. This is indeed the case. At a depth of 100 m at 28°W in Fig. 1, the maximum speed exceeds 80 cm s^{-1} when the winds are strong during the northern summer and autumn and

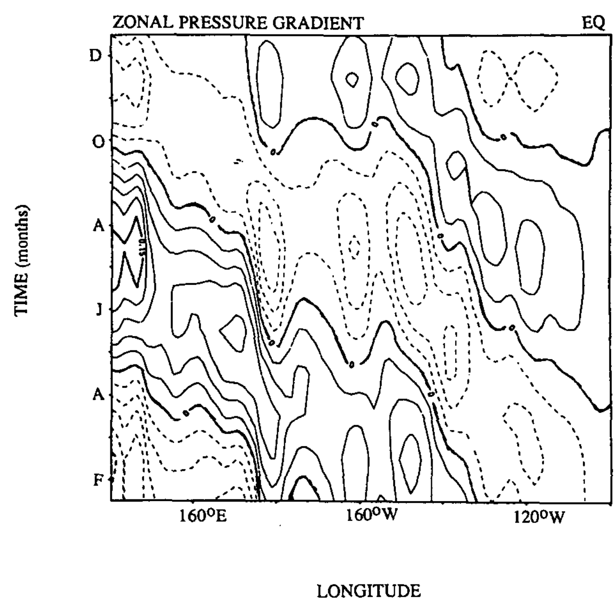

FIG. 6. The monthly departure of the vertically integrated zonal pressure gradient from the time-mean value in Fig. 5 as a function of month and longitude along the equator. The contour interval is 0.03 dyn cm^{-2}. Stippled contours correspond to values smaller than the mean.

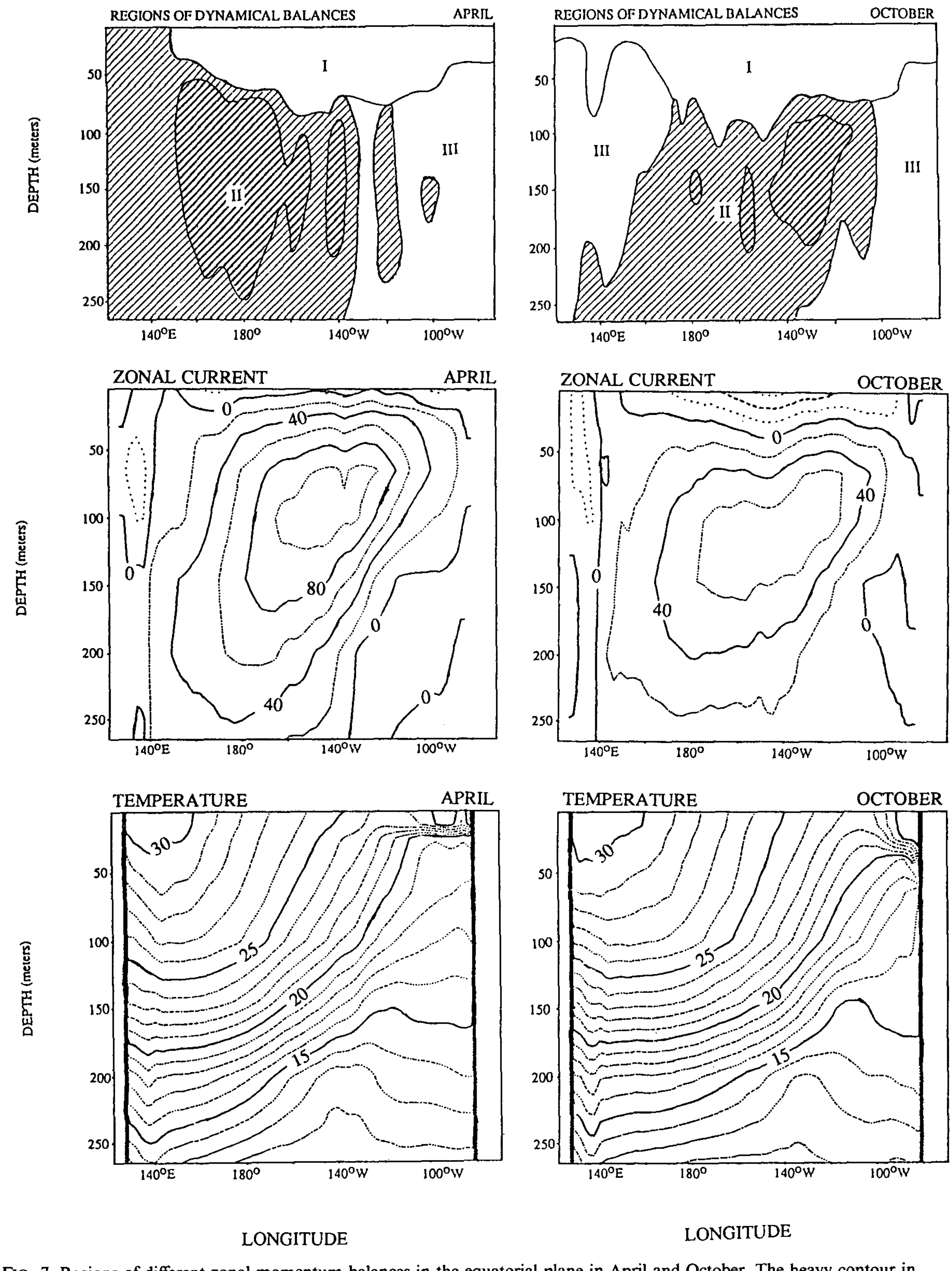

FIG. 7. Regions of different zonal momentum balances in the equatorial plane in April and October. The heavy contour in region II indicates where the acceleration exceeds 10^{-5} cm s^{-2}. The lower panels show the zonal currents and isotherms (°C) in the equatorial plane. For currents the contour interval is 20 cm s^{-1} and dashed contours indicate westward flow.

is less than 60 cm s^{-1} when the easterlies are weak during the early months of the year. To explain current variations closer to the surface is a more complicated matter.

In the core of the Equatorial Undercurrent, motion tends to be convergent so that a fluid parcel in the core tends to stay there. This is the reason why the Lagrangian approach adopted in the preceding discussion is appropriate. Motion in the surface layers of the equatorial regions is divergent so that the same approach is inappropriate. Although the momentum balance is given by Eq. (1) at all phases of the seasonal cycle, the surface flow is sometimes eastward, sometimes westward, as shown in Fig. 1. This indicates that the relative importance of the different terms in Eq. (1) changes with time, an inference confirmed by an analysis of the results. The driving force in the surface layers is the westward wind, which is represented by the term $(\nu U_z)_z$ in Eq. (1). The associated westward acceleration of the surface layers is countered principally by two sources of eastward momentum. One is the eastward pressure force, and the other is the upwelling of eastward momentum from the Equatorial Undercurrent. The balances change with time in both the Atlantic and Pacific in different ways for the two oceans.

In the surface layer at 0°, 140°W, the terms $(\nu U_z)_z$ and $(1/\rho_0)p_x$ wax and wane together as the winds intensify and weaken. The sum of these terms varies in the same manner, in contrast to the term WU_z, which has a relatively constant value because the shear U_z is large in the spring when the wind-induced upwelling W is weak, while the reverse is true in the summer and autumn. Thus, during the summer when the surface flow is westward, WU_z is relatively unimportant because the other terms are large; but in the spring when the surface flow is eastward, WU_z is an important source of eastward momentum.

In the surface layers of the western equatorial Atlantic neither the sum $(\nu U_z)_z + (1/\rho_0)p_x$ nor the nonlinear advection of zonal momentum varies much during the course of a year (Wacongne 1989). The surface flow tends to be weak and variable throughout the year. The exception occurs a short time after the intensification of the trades early in the northern summer—this can happen abruptly—when the magnitude of vertical mixing increases before that of the zonal pressure gradient does, so the surface flow can be westward at that time.

Finally, there is the question why vertical excursions of the equatorial thermocline are larger near 28°W than at 140°W. A given zonal wind stress determines the slope of the thermocline. If the wind fluctuates and causes the slope to change, then the associated variations in the depth of the thermocline are a strong function of position. It can be minimal in some longitudes that correspond to a nodal point and can be huge at other longitudes. Cane and Sarachik (1977, 1979) studied this horizontal redistribution of mass, associated with changes in the slope of the thermocline, for idealized winds. They find that thermocline displacements at a point depend strongly on its distance from the coasts. In reality matters are more complex because the winds vary spatially (Fig. 2), but the significant difference between 0°, 28°W and 0°, 140°W is that one is near a western coast in a region of intense winds while the other is essentially in midbasin.

4. Summary

To explain why the current fluctuations on the equator at 140°W in the Pacific and 28°W in the Atlantic are very different even though the local winds vary in a similar manner, it is necessary to take into account the winds farther west. In the equatorial Atlantic, conditions at 28°W are representative of conditions in the small oceanic region west of 28°W. In the Pacific where there is a huge region to the west of 140°W, the seasonal wind anomalies over the western Pacific can have a sign opposite to that of the anomalies in the east. In particular, when, during the northern spring, the winds east of 140°W are relaxed, the winds west of 140°W maintain an eastward pressure force that accelerates the Equatorial Undercurrent to the high speeds seen in Fig. 1. These high subsurface speeds mean that the upwelling of eastward momentum into the surface layers remains strong even though the intensity of local upwelling at 140°W decreases when the easterly winds weaken. This contributes to the eastward acceleration of the surface currents.

In the western equatorial Atlantic the seasonal change in the winds and in the slope of the thermocline also drive an undercurrent which, at a fixed depth of 100 m say, is intense when the wind is strong and weak when the wind is weak. Because of the small fetch of the accelerating pressure force, this change is modest in amplitude.

The explanations given here apply to simulations of the equatorial currents. The simulations have deficiencies that vitiate the validity of some of these arguments. Of particular concern are differences between the stratifications of the model and of the ocean. Whereas the western equatorial Pacific Ocean is observed to have a deep (approximately 100 m to 150 m) surface layer of uniform temperature, the upper layers of the model are stratified so that the thermocline tends to be more diffuse than in reality. The vertical structure of the pressure gradient, an important factor in the momentum balance of region I for example, could therefore be inaccurate. Although it is unlikely that the results presented here will change significantly, further studies with an improved model are necessary. Observational studies of the seasonal changes in the equatorial currents west of 140°W will also be valuable to check the validity of Fig. 7 for example.

Acknowledgments. Numerical modeling has played a significant part in the rapid progress of tropical oceanography over the past 20 years. The person who deserves the most credit for the development of this powerful tool, the general circulation model of the ocean, is Michael Cox. As coworkers at the Geophysical Fluid Dynamics Laboratory of NOAA in Princeton, we have had the benefit and pleasure of close contact with Michael Cox, whom we will always remember for his kindness and generosity.

During the course of this work one of us (Yi Chao) was supported by a grant from the NOAA program EPOCS and the other (S.G.H. Philander) by a NOAA TOGA Grant NA90AA-D-AC404.

REFERENCES

Bryan, K., 1969: A numerical method for the study of the world ocean. *J. Comput. Phys.*, **4**, 347–376.

Cane, M. A., and E. S. Sarachik, 1977: Forced baroclinic ocean motions II. *J. Mar. Res.*, **35**, 395–432.

——, and ——, 1979: Forced baroclinic ocean motions III. *J. Mar. Res.*, **37**, 355–398.

Chao, Y., 1990: Seasonal and interannual variability in the tropical Pacific Ocean. PhD Thesis, Princeton University, 131 pp.

Cox, M. D., 1984: A primitive equation, 3-dimensional model of the ocean. GFDL Tech. Rep. No. 1.

Halpern, D., and R. Weisberg, 1989: Upper ocean thermal and flow fields at 0°, 28°W (Atlantic) and 0°, 140°W (Pacific) during 1983–1985. *Deep-Sea Res.*, **36**, 407–418.

Horel, J. D., 1982: The annual cycle in the tropical Pacific atmosphere and ocean. *Mon. Weather Rev.*, **110**, 1863–1878.

Katz, E. J., R. Belevich, J. Bruce, J. Cochrane, W. Duing, P. Hisard, H. U. Lass, J. Meinke, A. de Mesquita, L. Miller and A. Rybnikov, 1977: Zonal pressure gradient along the equatorial Atlantic. *J. Mar. Res.*, **35**, 293–307.

Levitus, S., 1982: *Climatological Atlas of the World Oceans.* NOAA Prof. Paper No. 13, U.S. Govt. Printing Office, 173 pp., 17 microfiches.

Lucas, R., and E. Firing, 1985: The Rossby wave in the central Pacific Ocean. *J. Phys. Oceanogr.*, **15**, 55–67.

McPhaden, M., and B. Taft, 1988: Dynamics of seasonal and intraseasonal variability in the eastern equatorial Pacific. *J. Phys. Oceanogr.*, **18**, 1713–1732.

Meyers, G., 1979: Annual variation in the slope of the 14°C isotherm along the equator in the Pacific Ocean. *J. Phys. Oceanogr.*, **9**, 885–891.

Oort, A. H., Y. H. Pan, R. W. Reynolds and C. F. Ropelewski, 1987: Historical trends in the surface temperatures over the oceans based on the COADS. *Climate Dyn.*, **2**, 29–38.

Pacanowski, R., and S. G. H. Philander, 1981: Parameterization of vertical mixing in numerical models of tropical oceans. *J. Phys. Oceanogr.*, **11**, 1443–1451.

Philander, S. G. H., 1979: Variability of the tropical oceans. *Dyn. Atms. Oceans*, **3**, 191–208.

——, 1990: *El Niño, La Niña and The Southern Oscillation.* Academic Press, 293 pp.

——, W. J. Hurlin and A. D. Seigel, 1987: A model of the seasonal cycle in the tropical Pacific Ocean. *J. Phys. Oceanogr.*, **17**, 1986–2002.

Wacongne, S., 1989: Dynamical regimes of a fully nonlinear stratified model of the Atlantic Equatorial Undercurrent. *J. Geophys. Res.*, **94**, 4801–4815.

——, 1990: On the difference in strength between Atlantic and Pacific undercurrents. *J. Phys. Oceanogr.*, **20**, 792–799.