Initial Conditions for a General Circulation Model of Tropical Oceans

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ABSTRACT

A general circulation model of the tropical Pacific Ocean, which realistically simulates El Niño of 1982–83, has been used to determine how different initial conditions affect the model. Given arbitrary initial conditions (not in equilibrium with the wind) the model takes almost a year to return to a state in which the currents and density gradients are in equilibrium with the winds. Errors in the absolute value of the temperature persist far longer, however, indicating that accurate density data are essential initial conditions. If the correct density field is specified initially, but no information is provided about the currents, then the model recovers the currents within an inertial period, except for the eastern equatorial region. That region is affected by equatorial Kelvin waves which are excited because the model is initially in an unbalanced state. The currents associated with these waves are relatively modest and do not affect the density field significantly. Because of the large zonal scale of the thermal field in the tropical Pacific, three or four high resolution meridional density sections appear adequate for the initialization of the model. This result, however, takes into account neither the energetic waves, with a scale of 1000 km, that are associated with instabilities of the equatorial currents nor other high frequency fluctuations in the ocean.

1. Introduction

Low frequency phenomena in the tropics, the Southern Oscillation and El Niño, for example, involve interactions between the ocean and atmosphere (Bjerknes, 1966; Philander, 1983; Gill and Rasmusson, 1983). Coupled models of the ocean and atmosphere are therefore needed to explore the predictability of these phenomena. An intermediate step is the development of general circulation models of the tropical oceans that are capable of simulating sea surface temperature changes in response to prescribed heat and momentum fluxes at the ocean surface. Such a model of the tropical Pacific Ocean, forced with the surface winds observed there during 1982 and 1983, has recently been used in a realistic simulation of El Niño of 1982 and 1983 (Philander and Seigel, 1985). Although there were inaccuracies in the windfield, the model was remarkably successful in reproducing the major observed features of that El Niño: the expansion of the area of warm surface waters from the western into the eastern tropical Pacific during the second half of 1982, the simultaneous appearance of eastward surface currents at the equator, while the Equatorial Undercurrent decelerated and finally disappeared by the end of 1982, the return of the Undercurrent in early 1983 while warm surface waters persisted in the eastern tropical Pacific, and the termination of El Niño with the reappearance of cold surface waters, initially on the equator near 140°W in May 1983. The model is currently being used to simulate low frequency variability in the tropical Pacific over an extended period that includes several El Niño events and has been used in a realistic simulation of the seasonal cycle of the tropical Atlantic Ocean (Philander and Pacanowski, 1984).

An important reason for the success of the general circulation model (and of simpler models such as that of Busalacchi and O'Brien, 1981) in simulating low frequency variability of the tropical oceans is the relative unimportance of instabilities of the large-scale oceanic circulation of tropical oceans. Instabilities associated with the shear of the zonal currents near the equator can be energetic at times, but the resultant waves are remarkably linear and contribute to variability only in a narrow frequency range at a period of a month approximately (Philander et al., 1985). Variability at longer periods is induced primarily by the surface forcing and, in principle, can be simulated for an indefinite length of time provided the surface forcing is specified. Amplification of errors in initial conditions, a factor which severely limits the predictability of atmospheric motion in midlatitudes (Lorenz, 1984), is unimportant in models of the tropical oceans because the physical instabilities that cause the amplification of errors are important only at periods near a month and do not cascade energy over a wide range of frequencies.

An important factor that limits the ability of a general circulation ocean model to simulate low frequency variability in the tropics over extended periods, even when accurate surface boundary conditions are specified, is the ad hoc parameterization of mixing processes. Flaws in the parameterization cause the sharp shallow tropical thermocline to diffuse downward, thus affecting the thermal field in general and the sea surface temperatures in particular. The models are reasonably good at simulating changes in the topography of the thermocline on time scales comparable to the adjustment time of tropical oceans (of the order of a year), but the models are poor at simulating the processes that maintain the thermocline itself, on time scales of years and decades. (Simple one-level models take the existence of the thermocline for granted and simulate only changes in the topography of the specified thermocline.) It follows that if the general circulation models that are presently available were used operationally, then initial conditions will be required to restart the models at regular intervals. The parameterization of mixing processes is only one source of errors. Other sources of errors, inaccuracies in the surface boundary conditions and in the boundary conditions at high latitudes, for example, will also require that the model be restarted regularly. It is therefore important to know which, and how much, oceanographic data will be needed to restart a model.

This paper describes how different initial conditions affect the simulation of subsequent oceanic conditions in the tropical Pacific Ocean. The standard case, with which other cases are contrasted, is a simulation of conditions in the tropical Pacific Ocean from January 1982 onwards. (The simulation has become semi-operational because it continues at the beginning of each month when the mean winds for the previous month, obtained from the National Meteorological Center, Washington, DC, are used to advance the model another month. The results provide a coherent picture of current conditions in the tropical Pacific Ocean and are an aid in anticipating future developments in the Pacific.) Philander and Seigel describe results from the standard case for the period 1982 and 1983 when El Niño occurred. This paper compares results for the period that starts in March 1984 with results from other calculations in which the oceanic conditions at the beginning of March 1984 were changed. The surface winds remained unchanged. It has therefore been possible to determine how long it takes the model to return to the conditions of the standard case when different initial conditions are specified.

2. The model

The Pacific Ocean model extends from 130°E to 70°W and 28°S to 50°N. The coastlines and gridpoint distribution are shown in Fig. 1. The longitudinal resolution is a constant 100 km, but the latitudinal distance between grid points is 33 km between 10°S and 10°N and increases gradually poleward of this region. The spacing at 25°N is 200 km. The flat-bottom ocean is 4000 m deep. There are 27 levels in the vertical; the upper 100 m have a resolution of 10 m.

The primitive equations are solved numerically by means of finite differencing methods discussed in Bryan (1969). The use of Richardson number dependent vertical mixing coefficients is explained in detail in Pacanowski and Philander (1981). The coefficient of vertical eddy viscosity is assigned a constant value of 10 $cm^2 s^{-1}$ in the upper 10 m of the model to compensate for mixing by the high-frequency wind fluctuations which are absent from the monthly mean winds. The coefficient of horizontal eddy viscosity has the constant coefficient $2 \times 10^7 cm^2 s^{-1}$ equatorward of 10° latitude and elsewhere varies inversely as the latitudinal resolution to a value of $50 \times 10^7 cm^2 s^{-1}$ at 50°N.

Poleward of 20°S and 30°N the heat equation for the temperature T gains a term $\gamma(T - T^*)$, where T^* is the prescribed monthly mean climatological temperature for the region under consideration, and γ is a Newtonian cooling coefficient. Its value is 1/(2 days) near the zonal boundaries and decreases to a value of zero equatorward of 30°N and 20°S. This device mit-

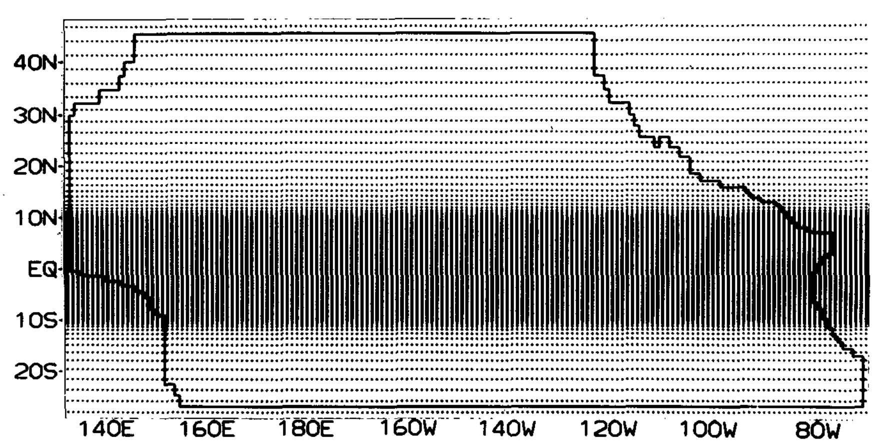

FIG. 1. Coastlines and grid-spacing of the Pacific Ocean model. Highest resolution is between 10°S and 10°N latitude.

igates the effect of the artificial zonal walls along the southern and northern boundaries of the ocean and forces the solution towards the climatology in these regions. There are similar terms in the horizontal momentum equations.

The heat flux across the ocean surface is

$$Q = \mathrm{SW} - \mathrm{LW} - \mathrm{QS} - \mathrm{QE}.$$

The solar short wave heating SW is taken to be 500 ly day^{-1} equatorward of 20° latitude and decreases linearly to 300 ly day^{-1} between 20° and 45° latitude. The long wave back radiation LW has the constant value of 115 ly day^{-1}. The sensible heat flux is

$$\mathrm{QS} = \rho C_D C_p V(T_0 - T_A),$$

and the evaporation is

$$\mathrm{QE} = \rho C_D L V[e_s(T_0) - \gamma e_s(T_A)](0.622/p_A)$$

where the saturation vapor temperature is

$$e_s(T) = 10^{(9.4-2353/T)}.$$

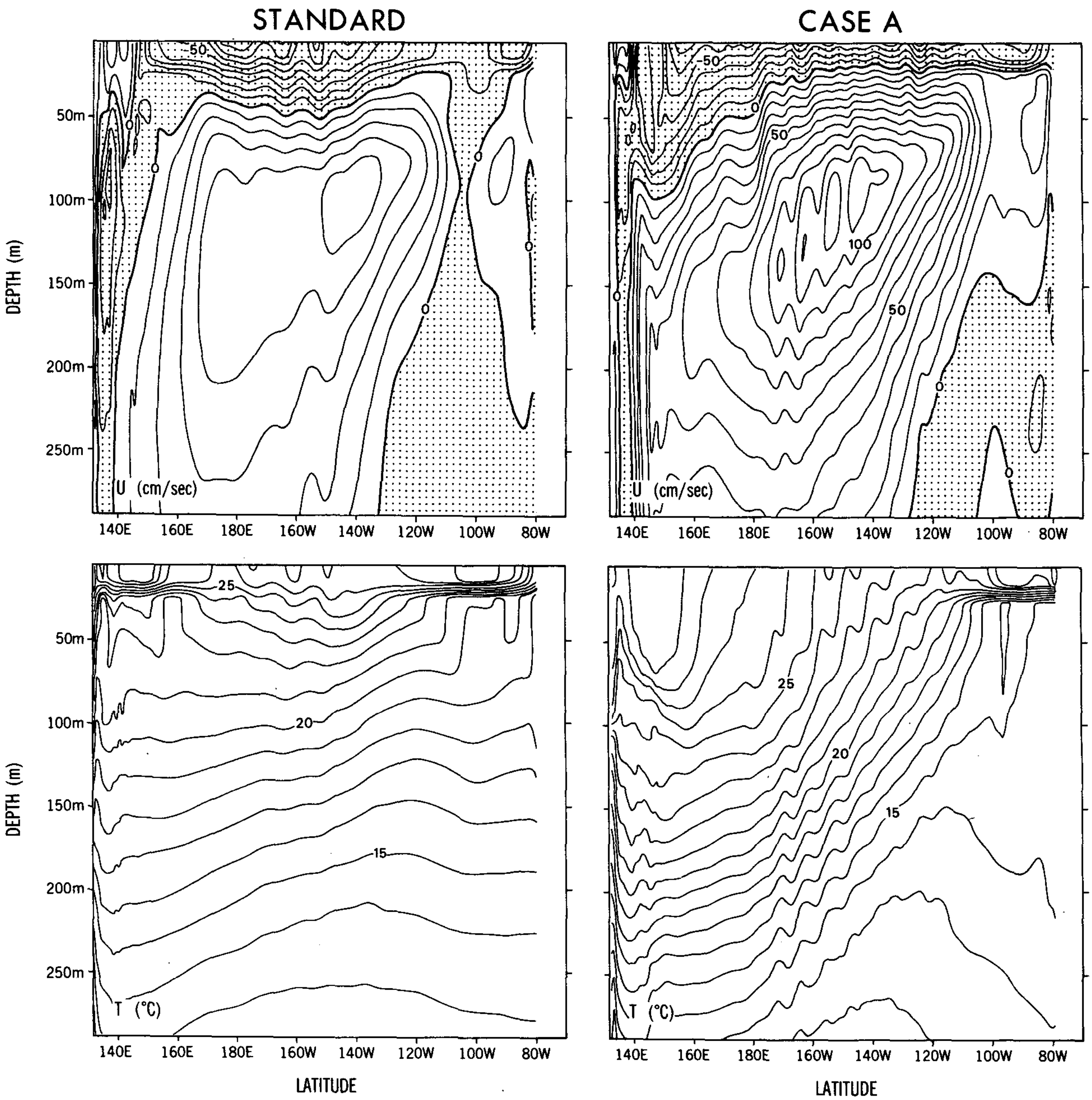

FIG. 2. Zonal sections of conditions on 15 March 1984 in the standard case (left) and Case A (right) at the equator. The top row is zonal velocity (in intervals of 10 cm s^{-1}) and the bottom row is temperature (in 1°C intervals). Shading indicates westward flow.

Here $\rho = 1.2 \times 10^{-3}$ g cm^{-3}; $L = 595$ cal g^{-1}; $C_D = 1.4 \times 10^3$; $p = 1013$ mb; $C_p = 0.24$ cal g^{-1} °C^{-1}; T_0 is the sea-surface temperature in degrees Kelvin; T_A is the atmospheric temperature at the surface; V is the surface wind speed; and the relative humidity γ is assigned the constant value 0.8. No provision is made for clouds. The sensible heat flux is found to be of secondary importance so that variations in heat flux are primarily a consequence of change in the evaporation. Since evaporation depends on the wind speed, avoidance of excessively high temperatures in regions of weak winds required that the wind speed not be less than 4.8 m s^{-1}. This minimum parameterizes evaporation caused by high-frequency wind fluctuations that are absent from the mean monthly winds.

3. The standard case

The initial conditions for the model are zero currents and the climatological temperature field (Levitus, 1982). Monthly averaged climatological winds then force the model for three years by which time the model

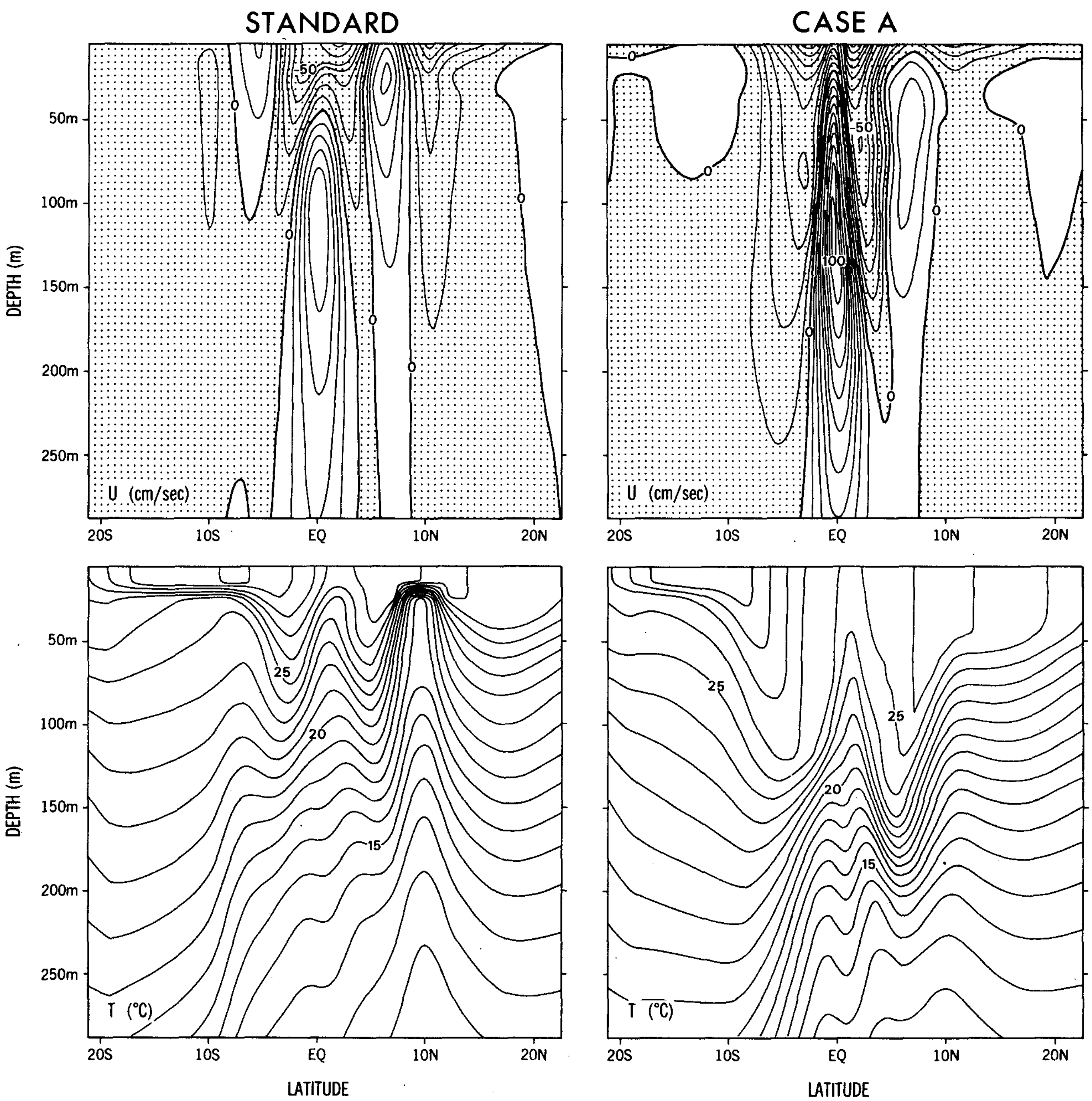

FIG. 3. As in Fig. 2 but of meridional sections along 159.5°W longitude.

has an equilibrium seasonal cycle. After this stage the model is forced with monthly mean winds provided by the National Meteorological Center. The atmospheric temperature T_A corresponds to climatological monthly mean surface temperatures except for the period 1982 and 1983 when $T_0 - T_A$ was specified to be 1°C. That part of the simulation that starts on 1 March 1984 is referred to as the standard case. In cases A and B, to be discussed later, conditions on 1 March are changed.

Figures 2 and 3 show the structure of the zonal flow and temperature along the equator and along a central meridian at the beginning of March 1984. The winds along the equator had been weak for the preceding six months so that the zonal slope of the thermocline is modest and the equatorial currents are weak. Along the central part of the basin there is both a North and South Equatorial Countercurrent. The thermocline has pronounced ridges near 9°N, 0° and 8°S.

Subsequent to March conditions evolved as shown in Figs. 4 and 5. Eastward surface flow appeared at the equator to the west of 150°W for a few months starting in June 1984. This coincided with a deepening of the thermocline in that region. To the east of 140°W westward surface flow persisted along the equator. Off the equator the eastward North Equatorial Countercurrent disappeared between May and July while the eastward South Equatorial Countercurrent decelerated.

In the next section we explore how the model returns to the conditions of the standard case when the initial conditions are altered.

4. Case A

In the absence of any measurements to initialize the model, it may seem practical to use climatological conditions as initial conditions. How will the model adjust if there happens to be a considerable difference between climatology and actual conditions? The standard case, just described in section 3, will be regarded as "actual conditions." In this section we replace those conditions on 1 March with climatological conditions (which are the conditions in the model when it is in equilibrium with the climatological winds). Figures 2 and 3 contrast the two sets of conditions. The equatorial currents are far more intense and the thermal gradients are far larger in the climatology than in the standard case. This is so because the climatological winds are much stronger than those that drive the model in the standard case. If the climatological conditions are used as initial conditions but the winds of the standard case are retained, then a vigorous oceanic adjustment can be anticipated because the ocean is far from a state of equilibrium initially. Figures 6 and 7 confirm this. At one stage the difference between the surface equatorial currents in this calculation and in the standard case are as high as 130 cm s^{-1}. Oceanic waves are excited to effect the adjustment back to a state of equilibrium.

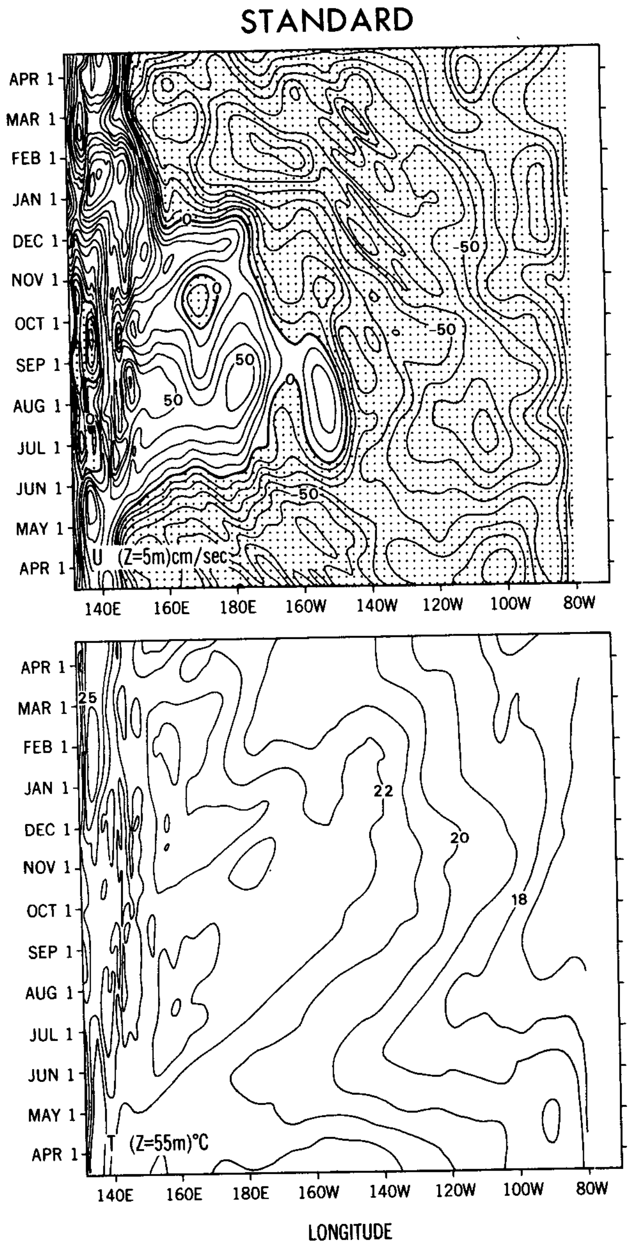

FIG. 4. Surface zonal velocity (contour interval = 10 cm s^{-1}) and 55 m temperature (contour interval = 1°C) fields at the equator from 15 March 1984 to 15 April 1985 for the standard case. Shaded areas are westward flow.

Equatorial Kelvin waves that reflect as Rossby waves at the eastern coast of the ocean basin are clearly evident in Fig. 6. The equatorial Kelvin waves also excite coastal Kelvin waves which propagate poleward and disperse into westward-traveling Rossby waves. Thus in Fig. 7, which shows variations along 6°N, a Rossby wave is seen to emanate from the eastern coast somewhat later than the Rossby wave that is excited at the coast on the equator. The small scale disturbances near the western coast in Fig. 6 are probably short Rossby

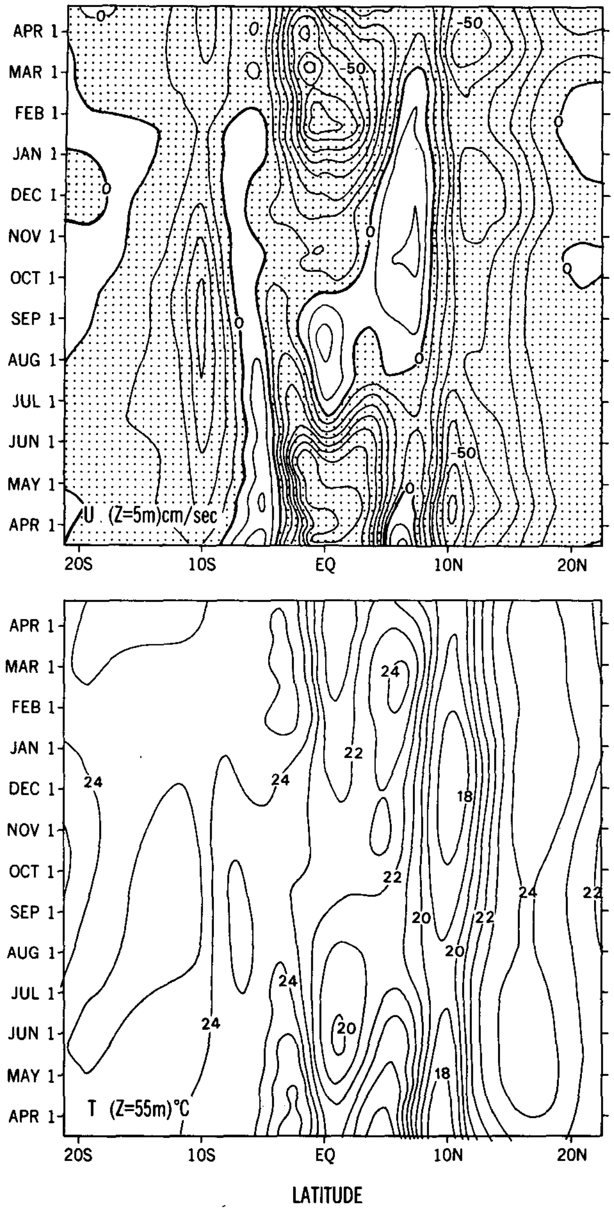

FIG. 5. As in Fig. 4 but along 155.5°W longitude from March 1984 to April 1985 for the standard case.

waves with eastward group velocities. Their influence is confined to a relatively narrow region near the western boundary of the basin.

Figures 6 and 7 show that it takes of the order of a year before the currents are essentially the same as those in the standard case. Below the thermocline, the oceans takes far longer to adjust (Philander and Pacanowski, 1980). In Fig. 8 it seems as if the adjustment time decreases with increasing latitude. This is not the case, however. The discrepancy between the initial and the standard currents happens to be large at the equator but relatively small at higher latitudes.

Although the currents start to approach those in the standard case within a year or so, it is clear from Fig. 6 that the adjustment time for the temperature is far longer. After a year the temperature at a depth of 55 m along the equator is still 4°C too high. Sections of the temperature and currents along the equator and along 155°W (Figs. 9 and 10) show that in spite of the absolute temperatures being too high, the horizontal temperature gradients are actually close to those of the standard case so that the currents are close to equilibrium. The reason for this was identified in the Intro-

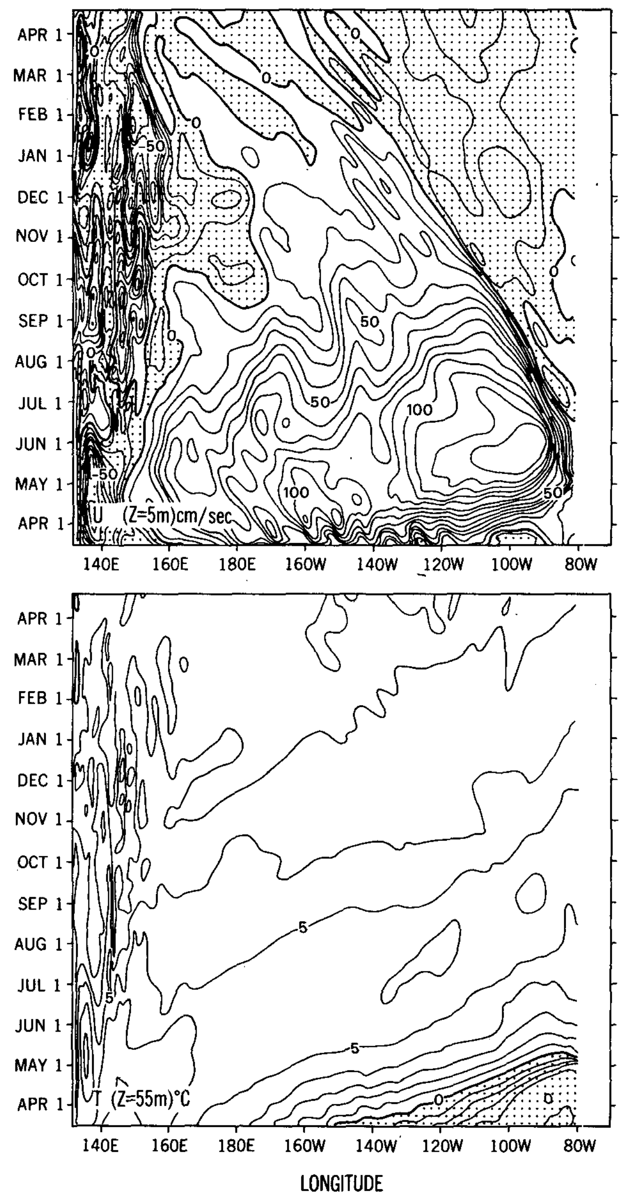

FIG. 6. The difference in surface zonal velocity (in 10 cm s^{-1} intervals) and 55 m temperature (in 1°C intervals) between the standard case, and Case A, along the equator. The difference is negative in shaded areas.

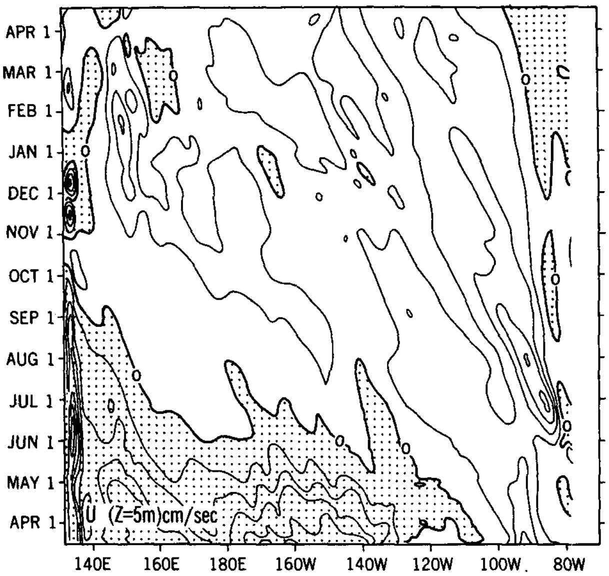

FIG. 7. As in Fig. 6 but at 5.83°N latitude from 15 March 1984 to 15 April 1985.

duction: the dynamical adjustment of the ocean is rapid but it affects only horizontal thermal gradients; the thermodynamic adjustment, which involves the surface heat flux, has a far longer time scale. If the initial thermal field is specified incorrectly then the model will rapidly generate the correct thermal gradients but the absolute temperature field will remain incorrect for a much longer period. If errors in the initial conditions are confined to one region then the dynamic adjustment, which involves a horizontal redistribution of mass in the upper ocean, will rapidly propagate the errors across the basin. It is clear that the initial conditions for the model must include accurate information about the thermal structure.

5. Case B

In reality, thermal data for the upper ocean are likely to be available along several sections, from expendable bathythermographs for example. Information about currents will be available at a few points only. On time scales longer than an inertial period oceanic currents are in geostrophic balance except possibly within a few hundred kilometers of the equator. It should therefore be sufficient to specify the density field in the ocean as the initial condition for the model without specifying the currents. The model will generate the currents that are consistent with the density field but the density field will change if this requires the conversion of a significant amount of potential energy into kinetic energy. The ratio of potential energy (P) to kinetic energy (K) for currents that are in geostrophic balance is

$$P/K = L^2/\lambda^2$$

where L is the horizontal scale of the currents and λ is the radius of deformation. If the horizontal scale of the motion is much larger than the radius of deformation—this is generally the case in the oceans—then the potential energy greatly exceeds the kinetic energy so that it is sufficient to specify the thermal field as the initial condition. The radius of deformation increases as the equator is approached and is comparable to the latitudinal scale of the flow (a few hundred kilometers) near the equator. It is therefore unclear whether it is sufficient to specify only the thermal field as initial conditions for a model of the equatorial oceans. To

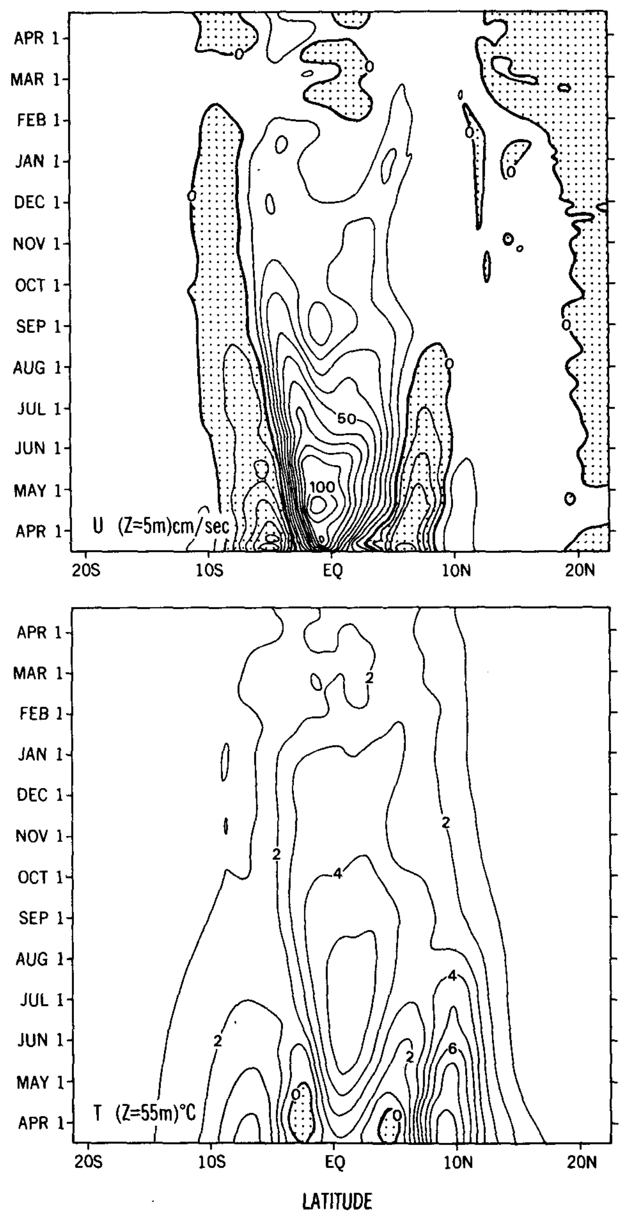

FIG. 8. As in Fig. 7 but of surface zonal velocity and 55 m temperatures along 155.5°W longitude.

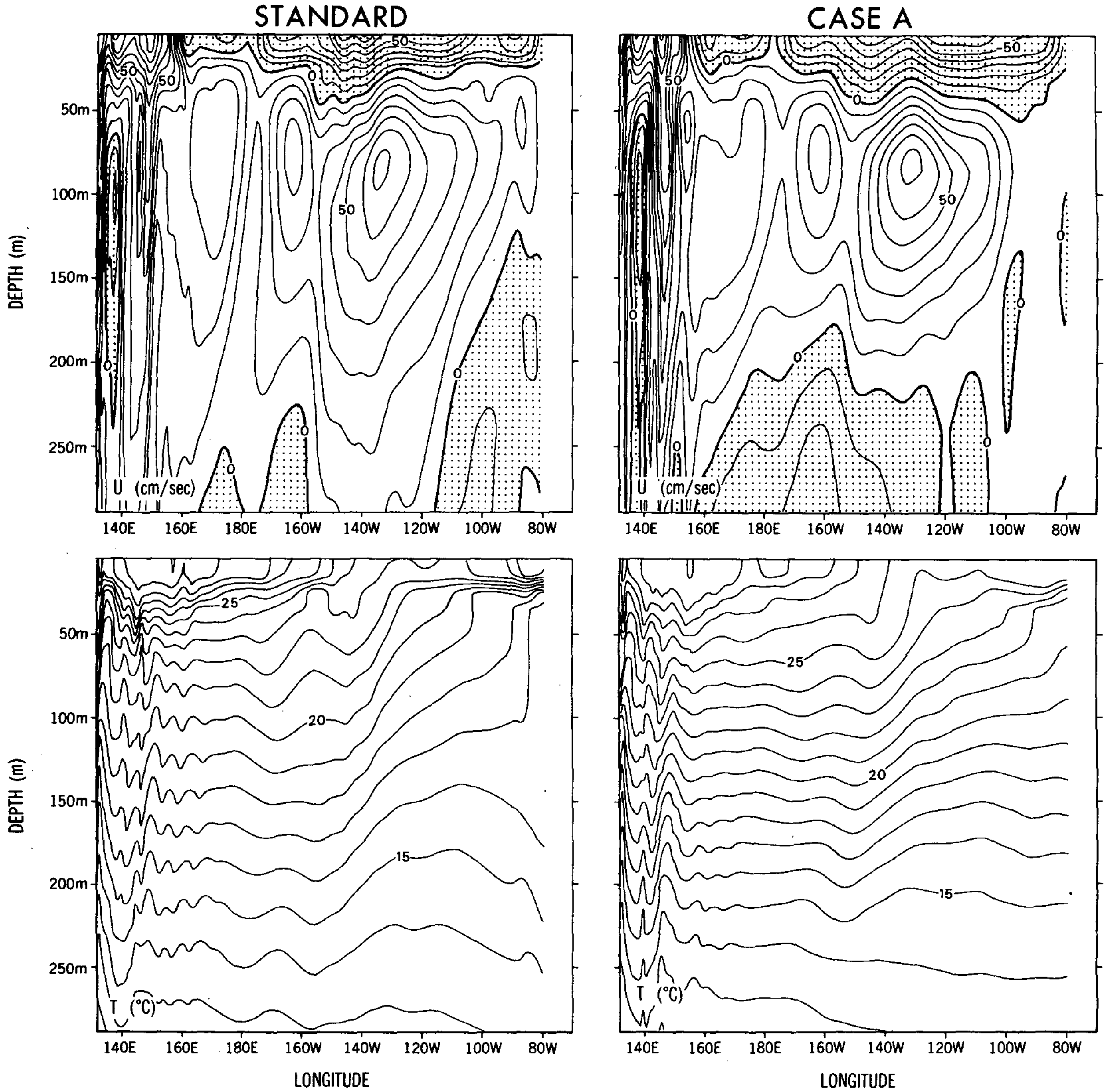

FIG. 9. Conditions on 15 December 1984 in the standard case (left) and in Case A (right) at the equator. The top row is zonal velocity and the bottom row is temperature. Contours and shading as in Fig. 6.

answer this question the standard case (section 3) was repeated from 1 March 1984 onward with the following changes in the initial conditions (Case B): all velocity components were set equal to zero; the temperatures above 400 m were retained but those below 400 m were replaced with climatological temperatures for March (from calculations in which the model is forced with climatological winds). These conditions were chosen because information about the thermal structure at depth will be sparse.

In the oceanic adjustment from the initially unbalanced state to an equilibrium state, waves are excited. Away from the equator these are primarily inertia–gravity waves. They are not evident in Fig. 11 which shows how the surface currents along 155°W were regenerated, because the data that are plotted are three-day averages. It is clear from Fig. 11 that, poleward of 4° latitude, the currents are established within an inertial period. Close to the equator not only inertia–gravity but also Kelvin and possibly Rossby waves are excited. The Kelvin waves travel eastward so that their cumulative effect is most pronounced in the eastern

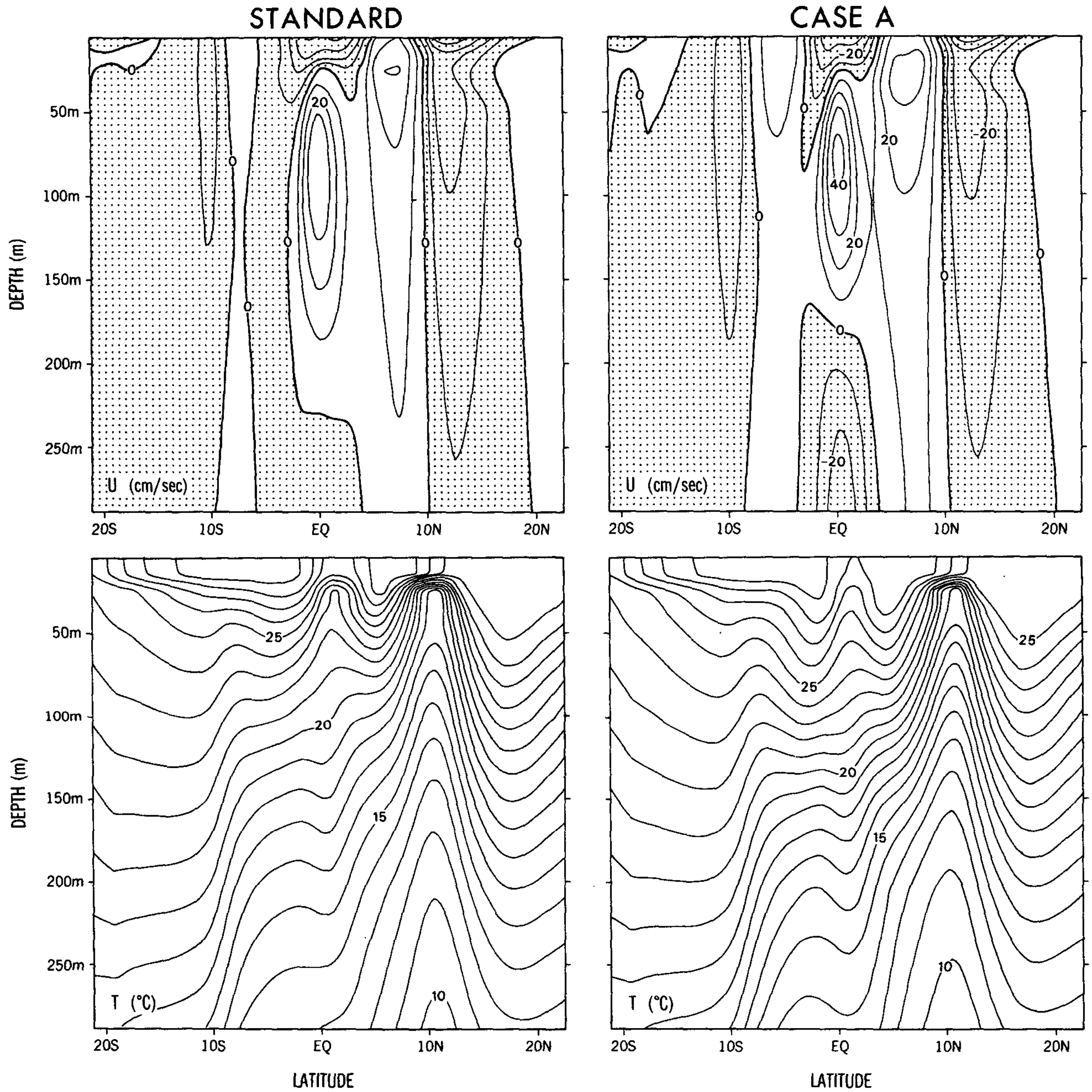

FIG. 10. Meridional sections of zonal velocity (in 10 cm s^{-1} intervals) and temperature (in 1°C intervals) on 15 December 1984 along 155°W. The left column is the standard case and the right column is Case A. Westward flow is shaded.

equatorial region. Figure 12 shows that the adjustment time for the equatorial region increases from west to east and is almost four months in the eastern side of the basin. Although there are anomalous currents of 20 cm s^{-1} for several weeks in Fig. 12, the temperature field never departed significantly from that in the standard case. (Near the equator where the Coriolis parameter vanishes, small latitudinal density gradient are associated with large currents.) Evidence for short Rossby waves in Fig. 12 can be found in the short disturbances near the western boundary. They seem unimportant.

The surface equatorial currents are to a large extent driven directly by the surface winds. The Equatorial Undercurrent, on the other hand, is driven by the pressure gradient which is maintained by the wind. This pressure gradient is present in the initial density field so that the Undercurrent evolved on the same time scale, and in a manner similar to that of the surface currents.

The conclusion from these calculations is that specification of the temperature field in the upper ocean is an adequate initial condition for the model except in

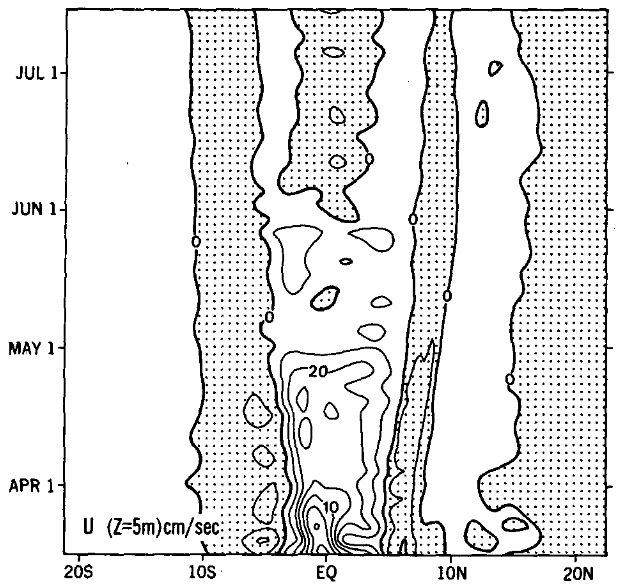

FIG. 11. Difference field (Case B minus the standard case) of surface zonal velocity along 155.5°W longitude from 15 March 1984 to 15 July 1984. The contour interval is 10 cm s^{-1}. Shaded areas are negative values (net westward flow).

the eastern equatorial region where it takes several months for equilibrium currents to be generated. There is no evidence that conditions below the thermocline affect the upper ocean. The adjustment of the deep ocean is much slower than that of the surface layers (Philander and Pacanowski, 1980) but this is of no concern since it is the upper ocean that is of principal interest.

6. Case C

Thermal measurements during a given month are likely to be available along a few sections only, not on the grid of Fig. 1. Initial data for the model will therefore have to be generated by interpolating between the sections. As a test we retained the temperature along the meridians 160°E, 160°W and 110°W on 1 March 1984 from the standard case and interpolated linearly between these meridians to generate data at all the grid points of the model. The results were practically indistinguishable from those described in section 5. The reason is the large zonal scale of the initial thermal field (Fig. 13). Although this large scale is a realistic feature, there are also energetic features with smaller scales, especially the 1000 km waves that are attributed to instabilities associated with the shear between the eastward North Equatorial Countercurrent and westward South Equatorial Current. These waves happened to be absent from the standard case on 1 March 1984. We return to this matter in section 7.

7. Discussion

The calculations described here indicate that the necessary initial conditions for a general circulation model of tropical oceans are accurate measurements of the thermal structure of the upper ocean. Thermal gradients, if initially incorrect, will take of the order of a year to come into equilibrium with the winds in the case of an ocean basin the size of the Pacific. The initial errors will, however, continue to be evident in the absolute value of the temperature for a period much longer than a year. This is clearly a problem if the model is needed to generate accurate sea surface temperatures.

Some of the calculations suggest that, because of the very large zonal scale of the thermal field in the tropics, the measurements to determine the initial conditions need not be dense in the zonal direction. Three or four meridional sections appear to be adequate. This result, however, does not take into consideration small scale variability such as the energetic 1000 km waves, near 3°N, in the central and eastern Pacific, that are caused by shear instabilities of the equatorial currents (Philander et al., 1985). These waves, fortunately, are linear and contribute to the energy spectrum in a narrow band of wavenumbers and frequencies (near a period of a month). It should therefore be possible to develop methods to filter out these waves from the data. There could still be problems with other fluctuations, that have relatively small scales. The matter needs further investigation.

Thermal measurements at depths greater than a few hundred meters are less critical than near-surface mea-

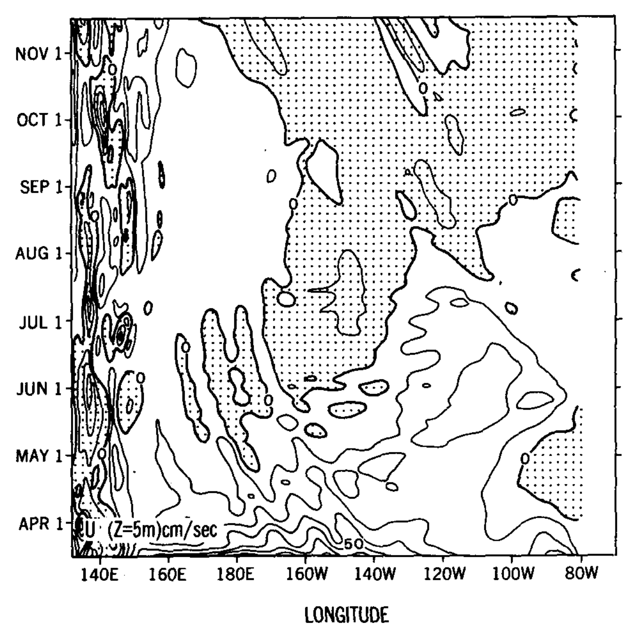

FIG. 12. As in Fig. 11 but along the equator from 15 March 1984 to 15 November 1984.

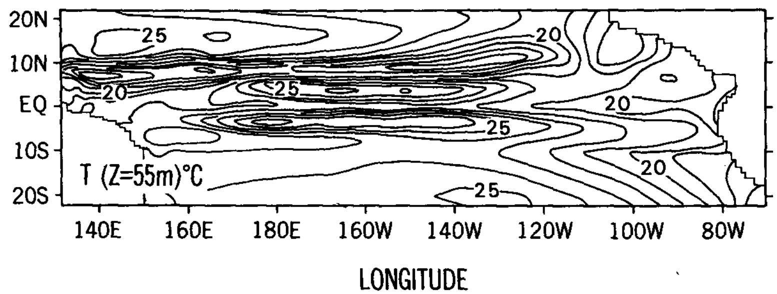

FIG. 13. Map of 55 m temperature (contour interval = 1°C) on 15 March 1984 for the standard case.

surements because there is no evidence that conditions at depth affect the upper ocean in the tropics.

Current measurements that are of most value for the initialization of the model are those in the upper equatorial ocean, especially in the eastern part of the basin. This is where the cumulative effect of equatorial Kelvin waves is most pronounced. Given accurate information about the density, but no information about the currents, it takes several months to generate the correct currents in the eastern equatorial region. The currents are modest—their speeds barely exceed 20 cm s^{-1}—and their effect on the thermal field is slight.

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REFERENCES

Bjerknes, J., 1966: A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. *Tellus,* **18,** 820–829.

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

Busalacchi, A. J., and J. J. O'Brien, 1980: The seasonal variability of the tropical Pacific. *J. Phys. Oceanogr.,* **10,** 1929–1952.

Gill, A. E., and E. M. Rasmusson, 1983: The 1982–1983 climate anomaly in the equatorial Pacific. *Nature,* **305,** 299–234.

Levitus, S., 1982: *Climatological Atlas of the World Ocean.* NOAA Prof. Paper 13, 173 pp., 17 microfiche, U.S. Govt. Printing Office, Washington, DC.

Lorenz, E. N., 1984: Irregularity: A fundamental property of the atmosphere. *Tellus,* **36A,** 98–111.

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., 1983: El Niño Southern Oscillation phenomena. *Nature,* **302,** 295–301.

——, and R. C. Pacanowski, 1980: The generation of equatorial currents. *J. Geophys. Res.,* **85,** 1123–1136.

——, and ——, 1984: Simulation of the seasonal cycle in the tropical Atlantic Ocean. *Geophys. Res. Lett.,* **II,** 802–804.

——, and A. D. Seigel, 1985: Simulation of El Niño of 1982–1983. *Coupled Ocean-Atmosphere Models,* J. Nihoul, Ed., Elsevier, 517–541.

——, D. Halpern, D. Hansen, R. Legeckis, L. Miller, C. Paul, R. Watts, R. Weisberg and M. Winbush, 1985: Long waves in the equatorial Pacific Ocean. *Eos, Trans. Amer. Geophys. Union,* **66,** p. 154.