Effect of Equatorial Currents on Surface Stress

R. C. PACANOWSKI

Geophysical Fluid Dynamics Laboratory/NOAA, Princeton University, Princeton, NJ 08542

23 July 1986 and 18 November 1986

The conventional computation of wind stress is done according to the relation:

$$\tau = \rho C_D|\mathbf{W}|\mathbf{W}, \quad (1)$$

where τ is the wind stress applied to the ocean, ρ is the density of air, C_D is the drag coefficient, and $\mathbf{W}$ is the velocity of the wind.

In shallow water ocean models the conventional approach is to apply this stress as a body force acting uniformly through the thickness of the ocean layer. In multi-level ocean models it is applied through the surface boundary condition $\tau = \nu\partial\mathbf{V}/\partial Z$ where ν is the viscosity at the ocean surface and $\mathbf{V}$ is the ocean current in the uppermost level. There have been several studies of the dependence of C_D on factors such as wind speed and air–sea temperature difference (Bunker, 1976; Garratt, 1977; Large and Pond, 1981). This note explores the effect of ocean currents on surface stress by writing:

$$\tau = \rho C_D|\mathbf{W}-\mathbf{V}|(\mathbf{W}-\mathbf{V}). \quad (2)$$

In equatorial regions, surface currents approach 1 m s^{-1} (Richardson and McKee 1984; Weisberg 1984) while surface wind speeds are around 6 m s^{-1} (Hellerman 1980). Since stress is dependent on relative motion between ocean and atmosphere, ocean currents can affect stress. Although usually neglected, the effect of this is quantified using a simulation of the seasonal cycle in the tropical Atlantic Ocean from a general circulation model forced with the climatological winds of Hellerman and Rosenstein (1983). For a description of the model and results see Philander and Pacanowski (1984, 1986, hereafter abbreviated PP).

Figure 1a shows the zonal component of the model velocity at a depth of 5 meters during mid-July from the above mentioned simulation; Fig. 1b shows the zonal stress. The corresponding result given by Eq. (2) is shown in Fig. 2a where the Hellerman stress has been modified by an amount shown in Fig. 2b (C_D was 1.2×10^{-3}). The most striking change in the simulation of the seasonal cycle is in the neighborhood of the equator where the speed of the westward surface current is decreased by approximately 30% when (2) is used. The explanation is that the current is in the direction of the wind so that the effective stress is decreased.

To the north of the equator the speed of the eastward North Equatorial Countercurrent is also reduced even though the direction is opposite to that of the wind. The effective westward wind stress is increased and this apparently slows the eastward current down. Whether surface currents flow against or in the same direction as the local wind the result is the same: a reduction in the speed of the surface current. It is noteworthy that Dewar (1984) obtained a similar result in his study of the effect of currents in Gulf Stream rings on wind stress.

Figure 3 shows a comparison of annual mean zonal velocity between the model simulation of PP and data from Richardson and Philander (1986). The model zonal velocities between 25° and 30°W at a depth of 5 meters were averaged for comparison with climatological ship drift data. Note that the model over estimates the extremes along with the shear between North Equatorial Countercurrent and South Equatorial Current. Assuming that Hellerman and Rosenstein (1983) climatological stresses are reasonable, one possible explanation for this discrepancy is that vertical mixing in the model is too weak. Another is that the modification of wind stress by ocean currents plays a role. When taking this latter effect into account the model is seen to be in closer agreement with measurement.

The change in the currents affects their latitudinal shear and this in turn affects waves which result when the shear leads to instabilities. The waves are evident in Fig. 1a (near the equator and 30°W) and absent in Fig. 2a. This difference is shown more clearly in Fig. 4, which is a plot of upwelling at a depth of 50 m in the western Equatorial Atlantic. The solid curve is from the simulation of PP and the dashed curve shows what happens when the effect of ocean currents on wind stress is taken into account. The fluctuations with periods of approximately one month are waves which are present from June onwards in one case but not in the

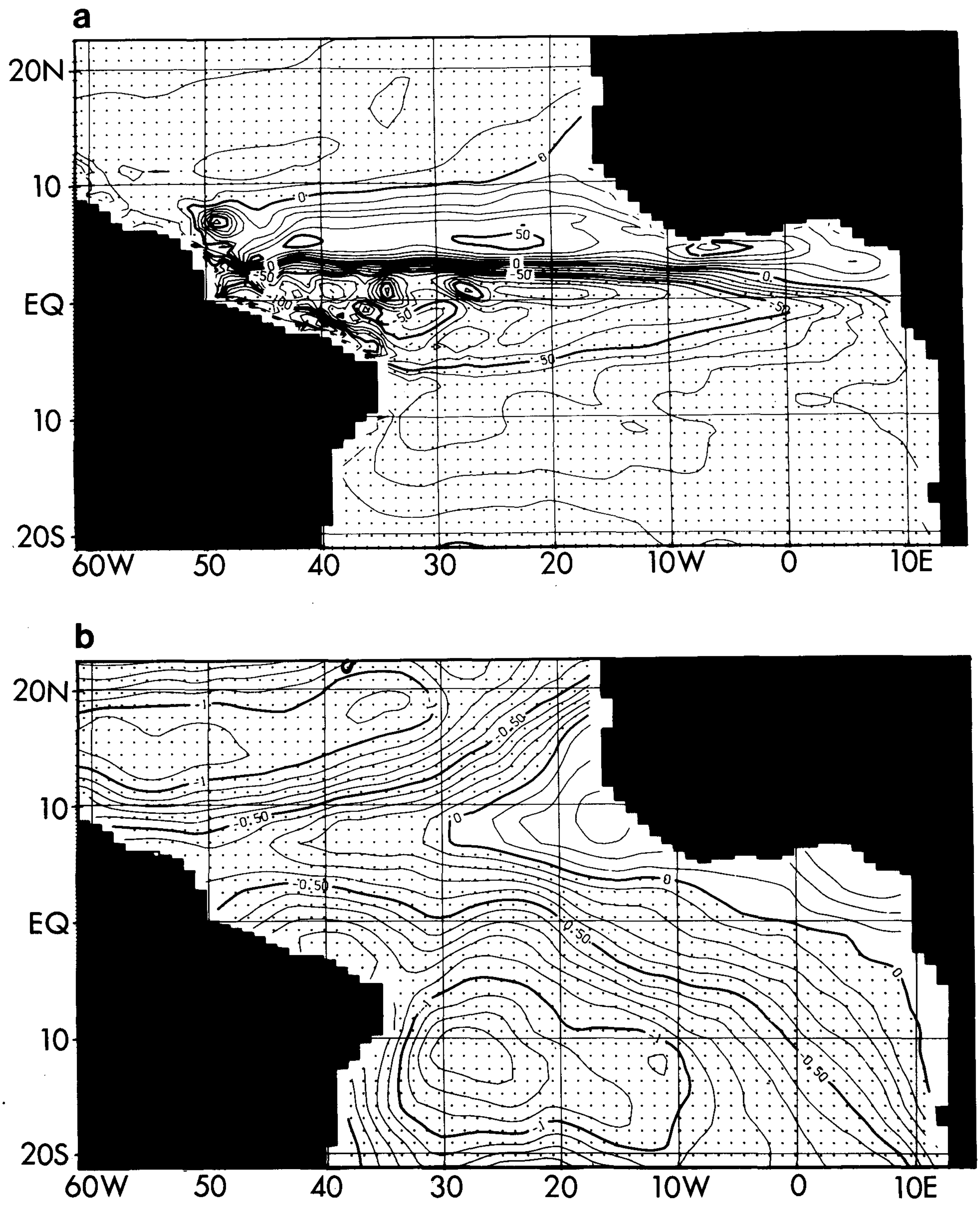

FIG. 1. (a) Zonal component of the velocity at 5 m depth during mid July from a simulation of the seasonal cycle in the Tropical Atlantic Ocean (PP). Shaded areas indicate westward flowing currents. Contour interval is 10 cm s^{-1}. (b) Zonal component of Hellerman and Rosenstein (1983) stress corresponding to the above result. Shading indicates westward stress and the contour interval is 0.1 dyn cm^{-1}/cm.

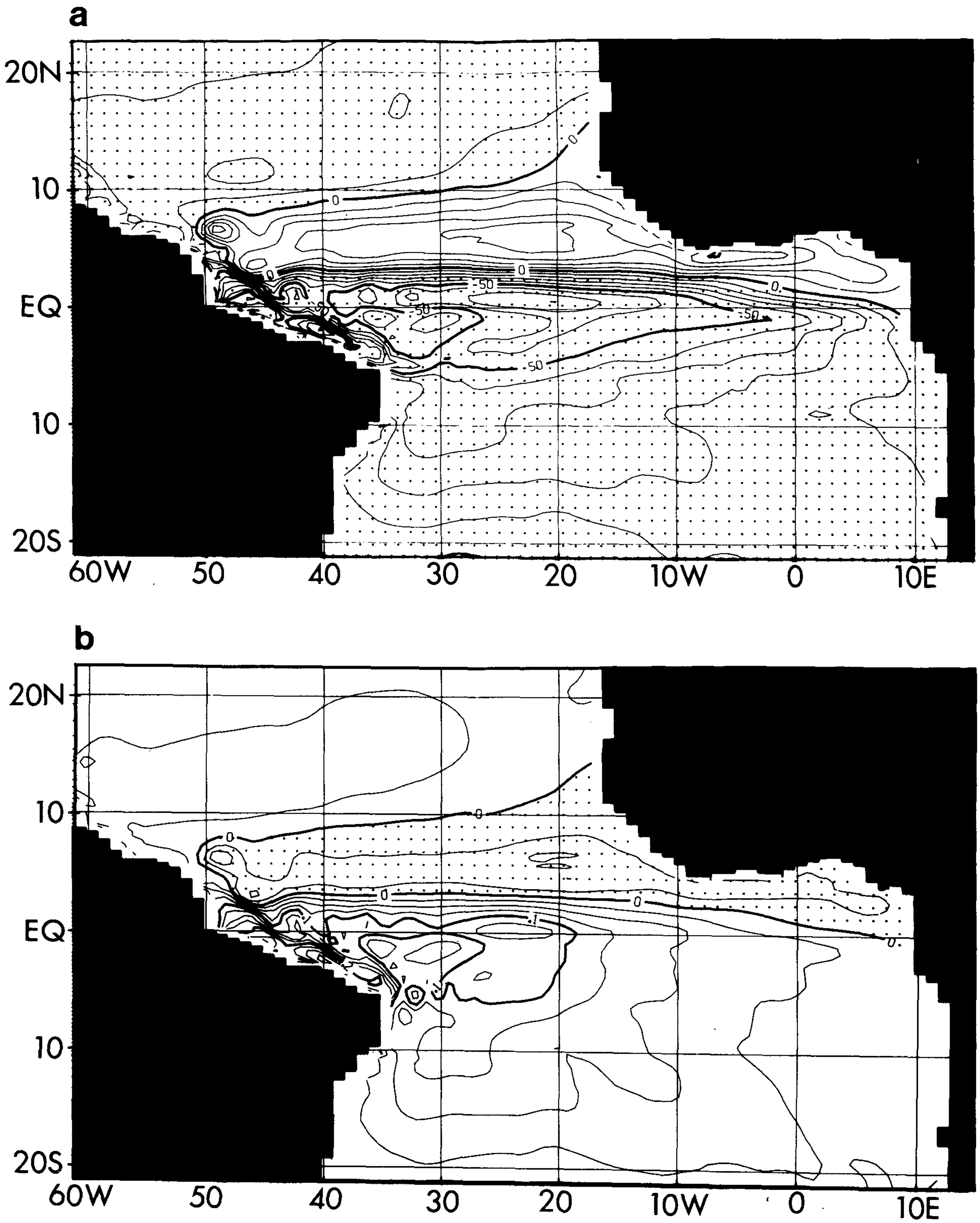

FIG. 2. (a) Zonal component of velocity as in Fig. 1 except taking into account the effect of ocean currents on wind stress. (b) Amount by which Hellerman and Rosenstein (1983) stress is modified when the effect of ocean currents is taken into account. Shading indicates a decrease(increase) in eastward(westward) stress. Contour interval is 0.025 dyn cm^{-1}/cm.

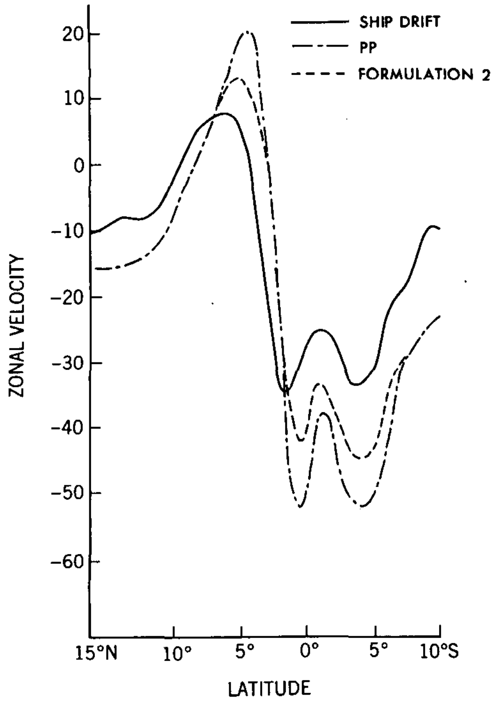

FIG. 3. Annual mean zonal velocity of profiles averaged between 25° and 30°W as a function of latitude in cm s^{-1}. Ship drift data is from Richardson and Philander (1986). PP is from the model simulation of PP. Formulation (2) is from taking the effect of ocean currents on wind stress into account.

other. Figure 4 also shows the extent to which equatorial upwelling is diminished when (2) is used. Since the wind is easterly near the equator (and in the same direction as the South Equatorial Current) the effective stress is reduced. This leads to reduced Ekman divergence and upwelling.

Of most meteorological interest is the change in SST. Comparison between Figs. 5a and 5b shows that Eq. (2) leads to a warming of 1 deg C in SST within the equatorial cold spot. The resulting differences [between stress Eqs. (1) and (2)] increase with increasing winds. Accurate SST is crucial in modeling air–sea interac-

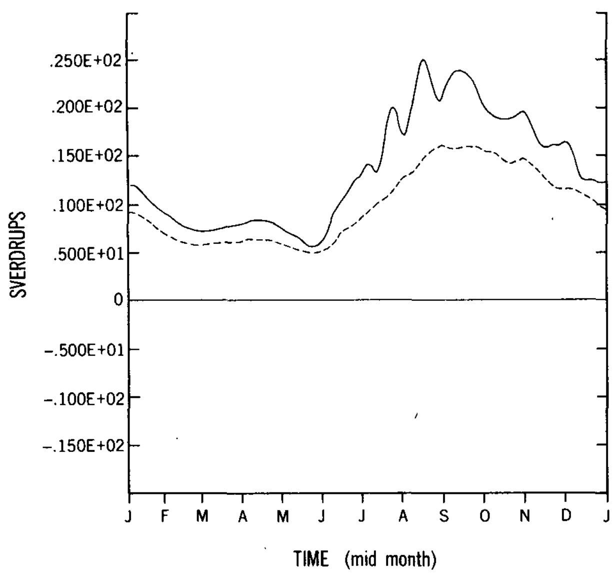

FIG. 4. Total upwelling at at depth of 50 m from 2.5°S to 2.5°N west of 30°W from a simulation of the seasonal cycle (solid curve) in the tropical Atlantic Ocean (PP). Dashed curve shows results when the effect of surface currents on wind stress is taken into account.

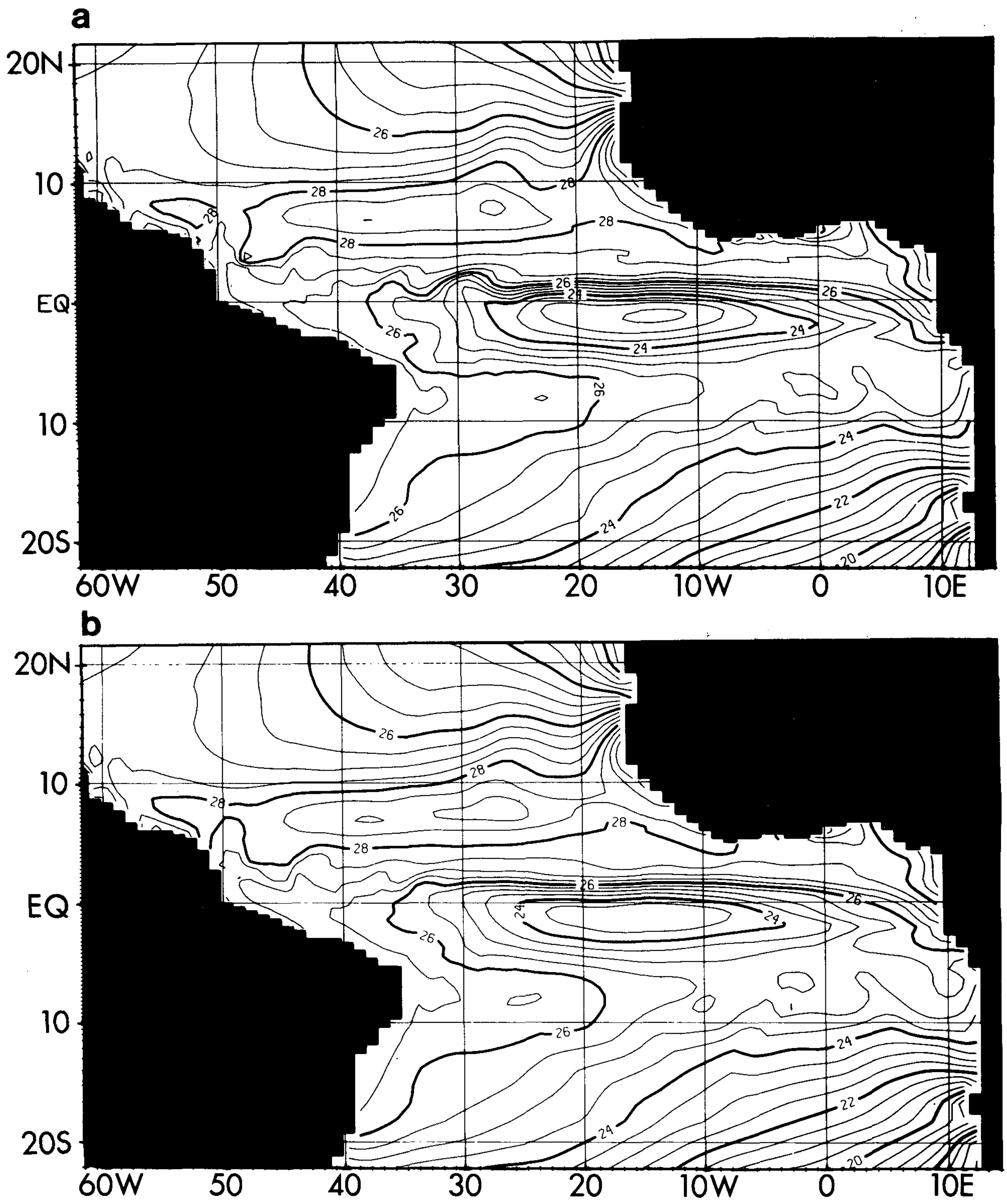

FIG. 5. (a) SST at a depth of 5 m during mid-July from a simulation of the seasonal cycle in the tropical Atlantic Ocean (PP). Contour interval is 0.5°C. (b) SST as in (5a) except taking into account the effect of ocean currents on wind stress.

tions. Given that Eq. (2) leads to the above mentioned change in SST using climatological winds, the effect on the modeling of air–sea interactions in the tropics should not be overlooked. This would be particularly relevant in the Equatorial Pacific where surface currents may exceed 1 m s^{-1}.

Previous studies of wind stress emphasize the sensitivity of stress to surface drag. It has been shown that in the model simulation of PP there is also a sensitivity to changes in wind stress due to ocean currents. When the final SEQUAL wind dataset is available the relative importance of these effects versus vertical mixing effects using daily and monthly wind fields will be investigated further.

REFERENCES

Bunker A. F., 1976: Computations of surface energy flux and annual air–sea interaction cycles of the North Atlantic. *Mon. Wea. Rev.*, **104,** 1122–1140.

Dewar W., 1984: Atmospheric interactions with Gulf Stream rings. Unpublished Ph.D. thesis. MIT–Woods Hole Joint Program, 230 pp.

Garratt J. R., 1977: Review of drag coefficients over oceans and continents. *Mon. Wea. Rev.*, **105,** 915–929.

Hellerman S., 1980: Charts of the variability of the wind stress over the tropical Atlantic. *Deep-Sea Res.*, **26**(Suppl. II), 63–75.

——, and Rosenstein M., 1983: Normal monthly wind stress over the world ocean with error estimates. *J. Phys. Oceanogr.*, **13,** 1093–1104.

Large W. G., and Pond S., 1981: Open-ocean momentum flux measurements in moderate to strong winds. *J. Phys. Oceanogr.*, **11,** 324–336.

Philander S. G. H., and Pacanowski R. C., 1984: Simulation of the seasonal cycle in the tropical Atlantic Ocean. *Geophys. Res. Lett.*, **II,** 802–804.

——, and ——, 1986: Simulation of the seasonal cycle of the tropical Atlantic Ocean. *J. Geophys. Res.*, (in press).

Richardson P. L., and McKee J. K., 1984: Average seasonal variation of the Atlantic North Equatorial Countercurrent from ship drift data. *J. Phys. Oceanogr.*, **14,** 1226–1238.

—— and Philander S. G. H., 1986: The seasonal variations of surface currents in the tropical Atlantic Ocean: A comparison of ship drift data with results from a general circulation model. *J. Geophys. Res.*, (in press).

Weisberg R. H., 1984: Seasonal adjustments in the equatorial Atlantic during 1983 as seen by surface moorings. *Geophys. Res. Lett.*, **II,** 733–735.