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# Properties of Long Equatorial Waves in Models of the Seasonal Cycle in the Tropical Atlantic and Pacific Oceans

S. G. H. PHILANDER, W. J. HURLIN, AND R. C. PACANOWSKI

*Geophysical Fluid Dynamics Laboratory, NOAA, Princeton University, Princeton, New Jersey*

In general circulation models of the seasonal cycle, westward propagating waves, with an approximate wavelength of 1000 km and period of 3 to 4 weeks, in the western equatorial Atlantic and eastern equatorial Pacific derive their energy from the kinetic and potential energy of the mean flow. There is intense downwelling in the cold crests of the wave and upwelling in the warm troughs. The local meridional heat flux associated with the waves is of the order of 100 W m$^{-2}$, but their contribution to the net heat transport across the equator is small. The waves are highly nonstationary in time and inhomogeneous in space.

## 1. Introduction

Westward traveling waves with a wavelength of 1000 km approximately and a period of the order of a month are clearly evident in satellite photographs of the temperature front near 3°N in the central and eastern Pacific Ocean [*Legeckis*, 1977]. Measurements with a variety of instruments (current meters and thermistor chains on moorings, drifter buoys, and inverted echo sounders) have established many properties of these waves and confirm that they are caused in part by an instability associated with the latitudinal shear of the zonal equatorial currents, especially the westward South Equatorial Current and the adjacent eastward North Equatorial Countercurrent [*Hansen and Paul*, 1984; *Miller et al.*, 1985; *Philander*, 1976; *Philander et al.*, 1985]. Theoretical stability analyses indicate that some of the energy of the waves is derived from the potential energy of the mean flow [*Cox*, 1980]. Waves similar to those in the Pacific have been observed in the Atlantic Ocean [*Weisberg*, 1984]. The waves are also present in general circulation models that simulate the seasonal cycles of the tropical Atlantic and Pacific oceans realistically. (The models are forced with the climatological monthly mean winds described by *Hellerman and Rosenstein* [1983]. For more details about the models and about the simulations, the reader is referred to *Philander and Pacanowski* [this issue]. This note briefly describes the properties of the waves in these simulations of the seasonal cycles in the tropical Atlantic and Pacific oceans and draws attention to the large fluctuations in the vertical velocity component associated with the waves.

## 2. Results

The principal currents in the tropical Atlantic and Pacific oceans (the eastward Equatorial Undercurrent and North Equatorial Countercurrent and the westward South Equatorial Current) are zonal and are subject to considerable seasonal variations in both oceans. The surface currents are at their weakest during the northern hemisphere spring and are most intense during the northern hemisphere summer and autumn, both in reality [*Wyrtki*, 1974; *Wyrtki et al.*, 1981; *Richardson and McKee*, 1984; *Katz and Garzoli*, 1982] and in the models (Figure 1). These seasonal changes in the latitudinal shear of the currents affect the amplitude, period, and wavelength of the waves which depend on the shear. The mean currents and their latitudinal shear also vary with longitude, and this too affects the waves which are nonstationary in time and inhomogeneous in space. In Figure 2 the waves are seen to be absent during the northern hemisphere spring and early summer and are also absent

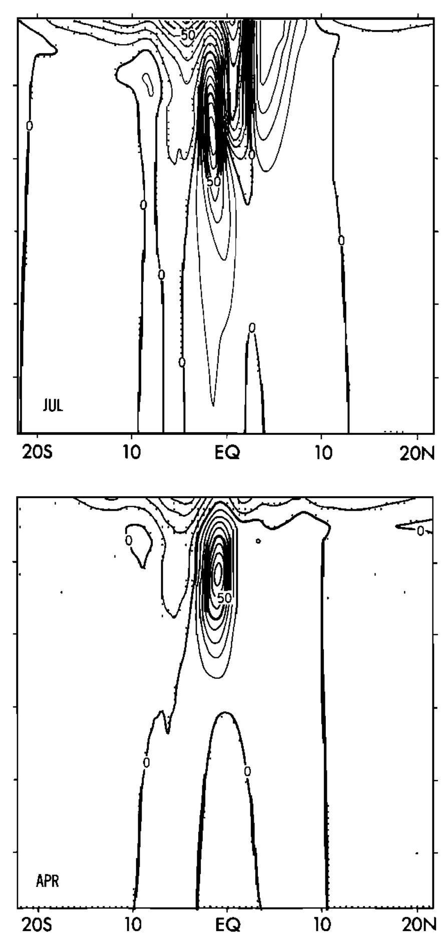


Fig. 1. The zonal velocity component across 30°W in the Atlantic Ocean on April 15 and July 15. The contour interval is 10 cm s$^{-1}$; flow is westward in shaded areas. The Pacific Ocean has similar seasonal changes in the mean currents.

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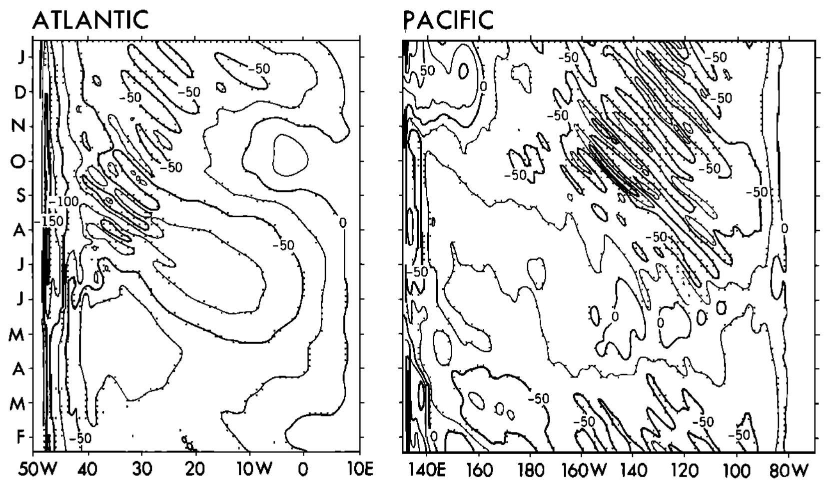


Fig. 2. Time dependence of the zonal surface current along the equator. The contour interval is 25 cm $s^{-1}$, and motion is westward in shaded areas.

from the eastern equatorial Atlantic and the far western equatorial Pacific. When the waves are present, their period is of the order of a month and their wavelengths is of the order of 1000 km, but these properties vary with time and longitude. The phase speed of the waves is always westward, but in the Atlantic they seem to have an eastward group velocity. The stability analyses performed thus far, which assume that mean conditions are uniform in time and longitude, do not explain the eastward group velocity [*Philander*, 1978; *Cox*, 1980; *Seigel*, 1985]. The seasonal intensification of the surface currents, which starts in early summer, is not uniform in longitude, so that the eastward group velocity may reflect an eastward progressing instability. Further analysis of the model results is needed to confirm this.

Figure 3 is an instantaneous picture of the sea surface temperature pattern, the current vectors in the surface layers, and the vertical component of the velocity. Motion is seen to be convergent in the cold crests of the waves where the northward flow is most intense. Motion is divergent in the warm troughs where the northward flow is weak. In other words, there is upwelling in part of the region where the sea surface temperature is high and downwelling where the sea surface temperature is low. The vertical speed at 60 m can vary between +350 cm $day^{-1}$ and −350 cm $day^{-1}$. At this depth, along 154°W, the annual mean vertical velocity in

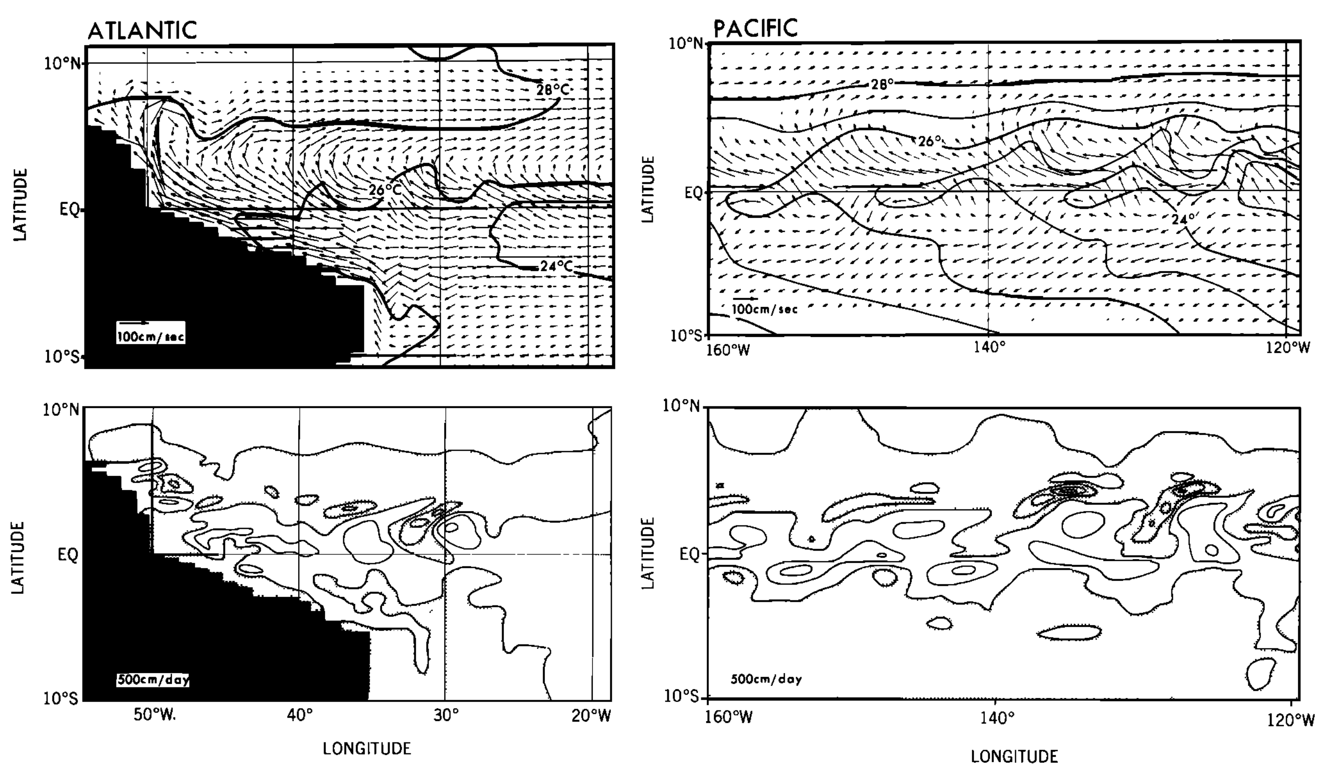


Fig. 3. (top) Surface current vectors, with surface isotherms superimposed, and (bottom) the vertical velocity component at a depth of 60 m on October 16. Motion is downward in shaded areas.

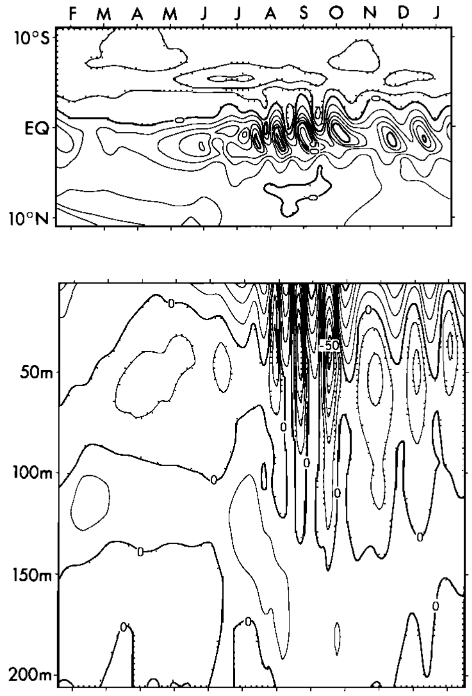


Fig. 4. (top) Time dependence of meridional surface flow along 30°W in the Atlantic Ocean, and (bottom) time dependence of the vertical structure of the meridional velocity component on the equator at 30°W in the Atlantic Ocean. Motion is southward in shaded areas. The contour interval is 10 cm $s^{-1}$. The meridional and vertical structures of the fluctuations in the Pacific Ocean are similar.

the seasonal model has a positive maximum (upwelling) of 450 cm $day^{-1}$ at the equator, and has a negative minimum (downwelling) of 200 cm/day near 3°N. The amplitude of the fluctuations is therefore comparable to that of the mean vertical velocity component.

The latitudinal region over which the waves have a large amplitude is confined to the region of large latitudinal shear and extends from 3°S to 5°N, approximately (Figure 4). In the vertical, the region of large latitudinal shear is confined to the upper 100 m, approximately. (It depends on the depth of the thermocline which varies with longitude.) This is also the vertical extent of the fluctuations associated with the waves (Figure 4). There is little change in phase with depth over the upper 100 m in Figure 4. The instability waves in the upper ocean excite free oceanic waves that propagate downward to greater depths so that vertical phase changes are evident below the thermocline [*Weisberg et al.*, 1979; *Philander et al.*, 1985]. The deep waves are predominantly Rossby gravity waves.

The correlations between the fluctuations of the different variables imply that the waves transport heat and momentum. To calculate the correlations, data from the model were band pass–filtered to isolate fluctuations with a period between 10 and 60 days. Figure 5 shows the time averaged correlations along 154°W in the central Pacific Ocean. (Results for the western equatorial Atlantic are qualitatively similar, although magnitudes are smaller because the waves are less energetic.) In Figure 5 correlations are large in a region with a small vertical (the upper 200 m of the ocean) and latitudinal extent (between 3°S and 5°N, approximately).

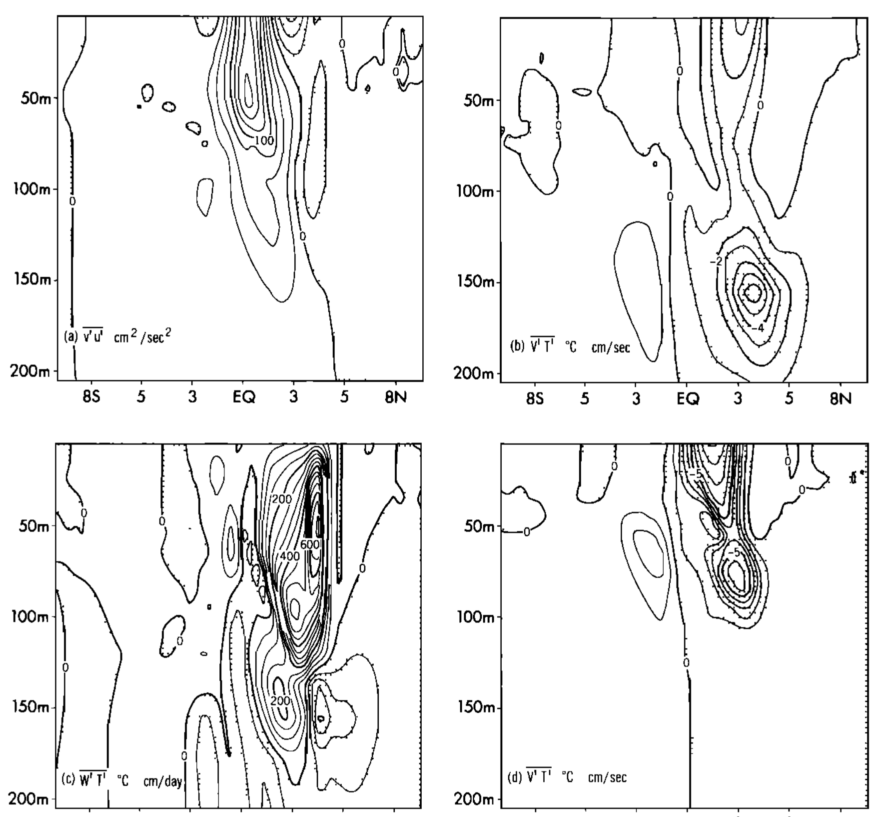


Fig. 5. Time-averaged correlations between (*a*) zonal and meridional velocity fluctuations ($\overline{u'v'}$), (*b*) between meridional velocity and temperature fluctuations ($\overline{v'T'}$), and (*c*) between vertical velocity and temperature fluctuations ($\overline{W'T'}$) along 154°W and (*d*) the correlations ($\overline{v'T'}$) along 110°W. Values are negative in shaded areas. The averages were taken over a 1-year period. The value of the coefficient of horizontal eddy viscosity was $2 \times 10^7$ $cm^2$ $s^{-1}$.

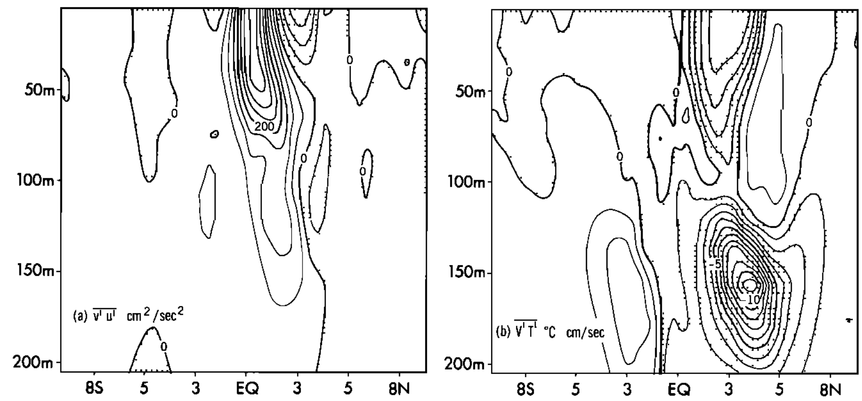


Fig. 6. Time-averaged correlations $\overline{u'v'}$ and $\overline{v'T'}$ along 154°W from a calculation in which the coefficient of horizontal eddy viscosity had the value of $1 \times 10^7$ cm² s$^{-1}$.

Near the surface the meridional structures of these correlations are in excellent agreement with those inferred from drifter buoy measurements by *Hansen and Paul* [1984]. The Reynolds stress term $\overline{u'v'}$ (where $u'$ and $v'$ denote zonal and meridional velocity fluctuations and the overbar indicates an average in time) is large and positive near the equator where the shear of the mean flow is negative, so that the eddies derive kinetic energy from the mean flow. The shear of the westward surface jet to the north of the equator appears to be most important, much more so that the shear between the Equatorial Undercurrent and this jet. The correlations between vertical velocity ($W'$) and temperature ($T'$) fluctuations $\overline{W'T'}$ are positive and indicate that the eddies also draw on the potential energy of the mean flow. This happens when cold water descends in the crests of the waves while warm water ascends in the warm troughs. Studies of the energetics of the waves, those by *Cox* [1980], and *Hansen and Paul* [1984], for example, usually assume that the mean flow is homogeneous in longitude and stationary in time. This is the case neither in the Atlantic nor in the Pacific Ocean. An analysis of the energetics that takes into account the spatial and temporal variations of the mean flow is a complex task and will be presented on another occasion. Figure 5 does, however, permit a comment on the heat flux effected by the eddies. The divergence near the surface of the correlation $\overline{v'T'}$ between the equator and 3°N implies an equatorward heat flux of the order of 100 W m$^{-2}$. Note that along 154°W this flux has a maximum at the depth of the thermocline (Figure 5*b*). The thermocline shoals to the east, so that the correlations $\overline{v'T'}$ are much more concentrated in surface layers along 110°W (Figure 5*d*). This heat flux by the eddies is comparable to that across the ocean surface. However, given the small depth over which the eddies are effective, their meridional heat transport is at most of the order of 10% of the estimated $10^{15}$-W northward annual mean heat flux across the equator in the Pacific and Atlantic oceans.

The results described thus far are from calculations in which the coefficient of horizontal momentum mixing was assigned the value $2 \times 10^7$ cm² s$^{-1}$. Figure 6 shows the correlations $\overline{u'v'}$ and $\overline{v'T'}$ from a simulation in which this eddy viscosity was assigned the value $1 \times 10^7$ cm² s$^{-1}$. The structure of the correlations is practically unchanged, but the magnitudes are larger. The time-averaged currents are essentially the same in the two simulations. In other words, the effect of the eddies is similar to that of horizontal diffusion. Numerical models with a resolution that is too coarse to resolve the eddies can simulate their effect by having a large coefficient of horizontal eddy viscosity.

In summary, waves associated with instabilities of the mean equatorial currents have very large signatures in all three velocity components and are nonstationary in time and inhomogeneous in space. Estimates of the heat and mass budgets (estimates of equatorial upwelling, for example) that are based on measurements (sections) that alias these waves may therefore not be reliable. Locally, heat fluxes by the waves are comparable to the fluxes across the ocean surface, but the waves make only a small contribution to the net meridional heat flux across a zonal plane in either the equatorial Pacific or the equatorial Atlantic Ocean.

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W. J. Hurlin, R. C. Pacanowski, and S. G. H. Philander, Geophysical Fluid Dynamics Laboratory, NOAA, Princeton University, P.O. Box 308, Princeton, NJ 08542.

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