Reprint from Journal of Marine Research, Volume 37, 1, 1979.

Upwelling in the Gulf of Guinea

by S. G. H. Philander[1]

ABSTRACT

Upwelling along the northern coast of the Gulf of Guinea occurs only between June and October even though the local winds are favorable for upwelling throughout the year and have no seasonal variability. Away from the coast, near the equator for example, the winds do vary seasonally and cause large scale oceanographic conditions in the Gulf of Guinea to change seasonally. One of these changes is an intensification of the eastward Guinea current north of the equator, during the summer. Because the current is in geostrophic balance it is associated with a thermocline that shoals in a northward direction, particularly so during the summer. Hence conditions along the northern coast are more favorable for upwelling during the summer than during the rest of the year. The cross-equatorial winds, that have a strong seasonal variation at the equator but not at the coast, are shown to contribute to the intensification of the Guinea current during the summer.

1. Introduction

Along the northern coast of the Gulf of Guinea the local winds are favorable for upwelling throughout the year. They blow parallel to the shore and cause an offshore Ekman drift which, between the Greenwich meridian and 8W, has practically no seasonal variability (Bakun, 1978). Yet low sea surface temperatures occur seasonally along this coast only between June and October. (See Fig. 1.)

It appears that the large-scale oceanographic conditions in the Gulf of Guinea from June until October are substantially different from the conditions during the rest of the year. During the nonupwelling season a surface layer of warm fresh water covers the eastern tropical Atlantic Ocean (Hisard and Merle, 1979). At this time the thermocline at the northern coast of the Gulf of Guinea is too deep for the local winds to induce upwelling. In June this warm surface layer disappears from most of the eastern tropical Atlantic so that the cold, saline subsurface waters are exposed. At the northern coast conditions are now favorable for upwelling. Associated with the disappearance of the warm surface waters during the boreal summer is an intensification of the eastward Guinea Current which is just north of the equator (Bakun, 1978; Hisard and Merle, 1979). (This current is an eastward extension of

1. Geophysical Fluid Dynamics Laboratory/NOAA, Princeton University, Princeton, New Jersey, 08540, U.S.A.

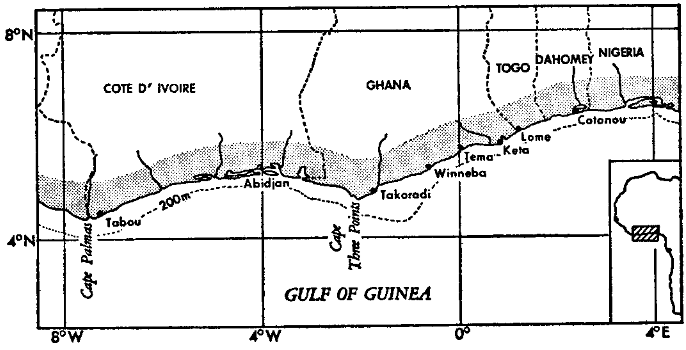


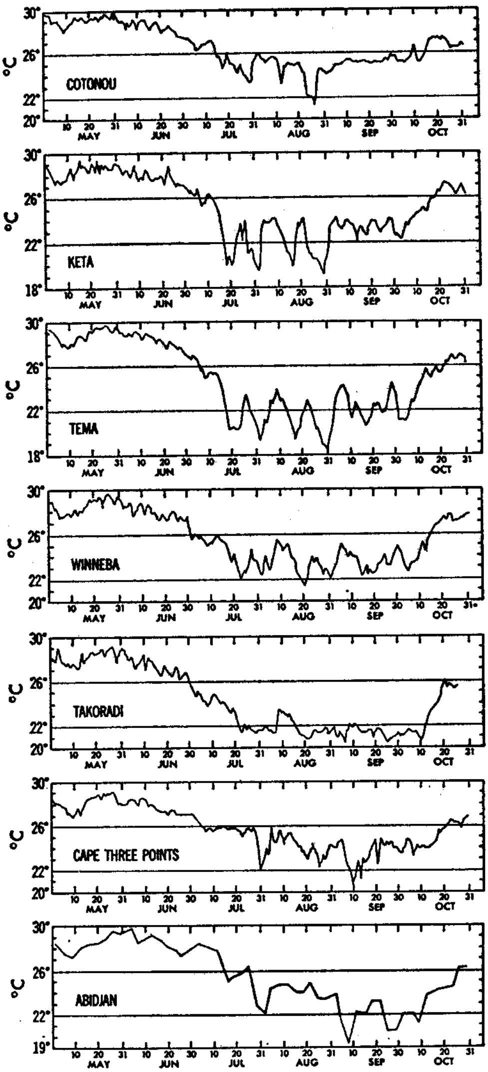
Figure 1. (a) Map of the Gulf of Guinea.

the North Equatorial Countercurrent.) The Guinea Current is in geostrophic balance and hence is associated with a latitudinal slope of the thermocline. Specifically the isotherms slope upwards toward the northern coast. An intensification of the current makes this slope more pronounced and results in a shallower thermocline near the coast. Hence an explanation for the intensification of the Guinea Current during the summer will contribute to an understanding of the coastal upwelling. (Cromwell (1953) first pointed out the significance of a geostrophically balanced current for upwelling.)

Zonal currents within 5° latitude of the equator can, to a considerable degree, be driven by cross-equatorial winds. In a linear model, in an unbounded ocean, northward winds drive an eastward jet north of the equator; it is centered on 3N approximately. The flow is antisymmetrical about the equator so that there is a similar westward jet in the southern hemisphere (Moore and Philander, 1977). Nonlinearities intensify the eastward jet significantly, and weaken the westward flow south of the equator (Cane, 1978). In response to the sudden onset of cross-equatorial winds these currents are in geostropic balance, and in equilibrium with the winds, after about one week. Hence winds that vary on a time-scale long compared to a week, drive variable currents that are always in equilibrium with the winds. If the winds have no spatial structure then the flow is essentially longitude-independent except near north-south coasts where boundary layers close the circulation (Cane, 1978).

These results are of relevance to the Gulf of Guinea because there the cross-equatorial winds are particularly intense during the summer, at and south of the equator. (See Fig. 2.) The results described above, however, are for an unbounded ocean and for winds with no spatial structure. How does a coast along 5N modify the results? Will the thermocline be shallow at the coast even if the winds are intense at the equator but not at the coast?

In this note we attempt to answer these questions by describing the response of a



(b) Sea surface temperatures as recorded daily at a number of coastal stations. (Courtesy of R. Houghton and J. Verstraete.) The pronounced 14-day oscillations are tidal in origin (Picaut and Verstraete, 1979) and are distinct from the upwelling phenomenon. From October until May the well-mixed surface layer masks these waves which are nonetheless present. The sudden drop in temperature at some stations in June is associated with these waves.

two-dimensional model—longitudinal variations are neglected—to the wind-stress shown in Figure 3. The model is described in Section 2; results are presented in Section 3 and discussed in Section 4.

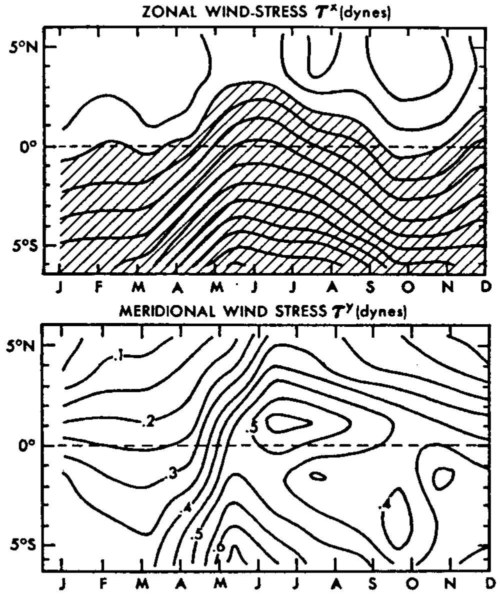


Figure 2. Monthly mean zonal (τ^x) and meridional (τ^y) wind-stress components along 4W. The shaded area indicates westward stress. The contour interval is .05 dynes. After Bunker (1977).

2. The model

Let (x,y,z) be the eastward, northward (from the equator) and upward (from the ocean surface) coordinates, and let u,v and w be the corresponding velocity components. We assume that the flow is independent of x so that a streamfunction can be introduced:

$$v = \psi_z \qquad w = -\psi_y$$

If we make the hydrostatic approximation then the vorticity for motion in the y-z plane is

$$\zeta = \psi_{zz}$$

The equations of motion of a Boussinesq fluid can now be written as follows:

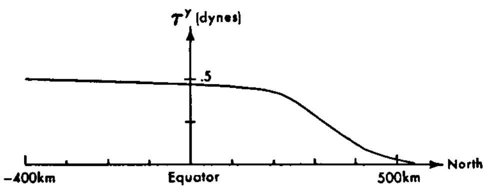


Figure 3. The latitudinal structure of the northward wind that drives the motion in the model.

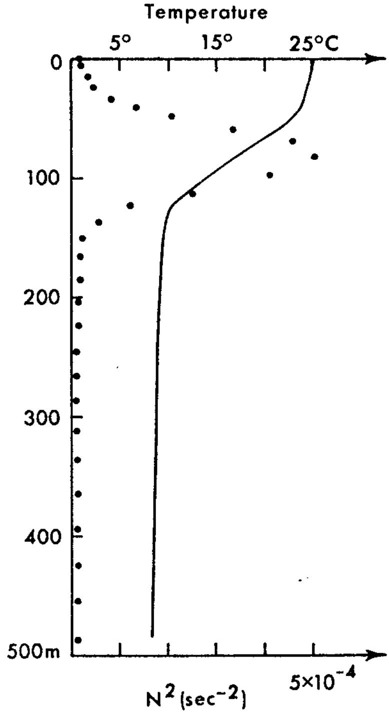


Figure 4. Initial temperature field and the associated Brunt-Vaisala frequency N in the upper layers of the model. The dots indicate the vertical distribution of grid points.

$$\zeta_t - J(\psi,\zeta) + fu_z = \frac{g}{\rho_0}\rho_y + \nabla \cdot (\nu\nabla\zeta) \qquad 1(a)$$

$$u_t - J(\psi,u) - f\psi_z = \nabla \cdot (\nu\nabla u) \qquad 1(b)$$

$$\rho_t - J(\psi,\rho) = \nabla \cdot (K\nabla\rho) \qquad 1(c)$$

Here $f = \beta y$ is the Coriolis parameter, g is the gravitational acceleration, t measures time, J denotes a Jacobian, ν and K represent coefficients of eddy diffusion, and ρ is the density.

The appropriate boundary conditions are

$$\nu u_z = 0;\ \nu\zeta = \tau;\ \psi = 0;\ \rho_z = 0 \text{ at } z = 0 \qquad 2(a)$$

$$u_z = 0;\ \zeta = 0;\ \psi = 0;\ \rho_z = 0 \text{ at } z = -H \qquad 2(b)$$

At vertical boundaries (along circles of latitude) we impose the conditions

$$\psi = u_y = \rho_y = \zeta = 0 \qquad 2(c)$$

These equations have been solved numerically by using a computer-program developed by Orlanski and Ross (1973) and Orlanski, Ross and Polinsky (1974). A centered space and time difference approximation is used to represent the space and time derivatives. The Jacobians are treated by the methods of Arakawa (1966) and Lilly (1965) to minimize nonlinear instability. The leap-frog method is used for time differencing but with the diffusive terms lagged one time step. The solution is time-smoothed every 30 time steps to minimize mode splitting.

The depth of the ocean is taken to be 3000m and the 61 levels in the vertical are spaced nonuniformly. (See Fig. 4) The ocean is bounded by a wall 520 km to the north of the equator and another 2000 km to the south of the equator. With 51 grid points the latitudinal resolution is 40 km. The grid is staggered. (Different variables are defined at different grid points so as to reduce finite differencing errors.) Initially the ocean is motionless and the density field ρ is a function of depth only. It is assumed that

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

where $\alpha = .0002°C^{-1}$. Figure 4 shows the initial temperature field and the associated Brunt-Väisälä frequency N. (Below a depth of 500 m, N has a constant value and T decreases linearly to zero at the ocean floor.)

The coefficient of vertical eddy viscosity ν_v has a value equal to 20 cm²/sec at the first 5 grid points (i.e. in a 35 m deep "mixed" layer) and has a value equal to 1 cm²/sec at greater depths. The coefficient of horizontal eddy viscosity ν_H has a value equal to 1000 ν_v. We assume that $K_v = 1$ cm²/sec and $K_H = \nu_H$. In some experiments (not described here) we assumed $\nu_v = 20$ cm²/sec everywhere or we assumed a more complicated function of depth for ν_v but such changes appeared to have little effect on the results for the first three weeks of integration.

Because the fluid is diffusive, the initial temperature field changes even in the absence of any forcing. (Temperatures decrease uniformly above the thermocline and increase uniformly below the thermocline.) For the chosen value of K_v these changes are minimal. Calculations for different values of K_v show that the velocity field and the latitudinal structure of the density field are relatively insensitive to the value of K_v.

3. Results

The initial stratification has been specified to be representative of the observed one during the winter: there is a nearly mixed layer to a depth of about 35 m and below that is a strong thermocline. (See Fig. 4.) This stratification is such that the equivalent depths also coincide with those of the equatorial Atlantic. The values for the first few baroclinic modes, and the associated radii of deformation, are (71 cm, 365 km), (27 cm, 286 km) and (14.5 cm, 245 km). The region of distinct equatorial dynamics will therefore extend a distance of the order of 500 km from the equator.

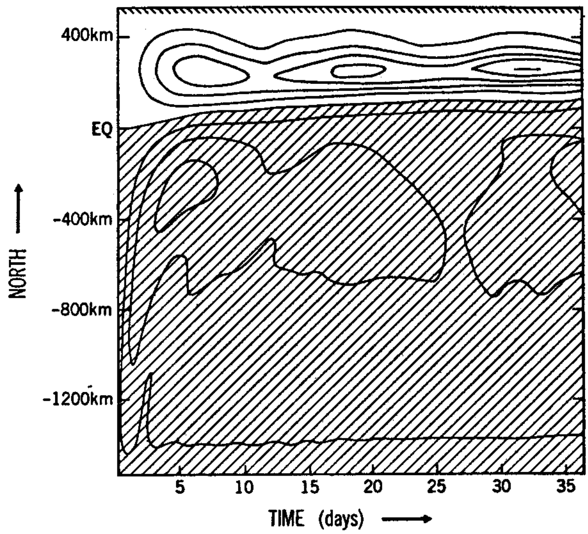


Figure 5. The development of the zonal velocity component at the surface as a function of time. Isopleths are at 10 cm/sec intervals. The shaded area indicates westward flow.

Extra-equatorial dynamics is valid poleward of approximately 5° latitude. Initially there is of course nothing distinctive about the equatorial region. For the values of the parameters given above, equatorial dynamics is important on a time-scale longer than about 5 days. The numerical calculations show that these are indeed the important spatial and temporal scales.

Figure 5 shows the evolution of the zonal velocity component at the surface. Initially the flow is westward south of the equator and eastward north of the equator but nonlinear advection, driven by the northward winds, rapidly causes the westward flow to penetrate into the northern hemisphere. After an adjustment period of a few days, the flow is nearly steady except for equatorially trapped inertia-gravity oscillations that were generated by the sudden onset of the winds. Figure 6a depicts the vertical structure of the zonal velocity component after 3 weeks. The eastward current is much more intense than the westward current, a consequence of nonlinearities which also cause a strong shear zone just north of the equator. The stream function for flow in a meridional plane (Fig. 6b) shows a closed cell confined to the surface layers and to the neighborhood of the equator: downwelling occurs in a narrow zone near 3N while upwelling takes place over a region of larger latitudinal extent, south of the equator. Near the coast the flow is not strongly divergent but in a meridional plane, the isotherms near the northern coast nonetheless slope upwards toward the surface (Fig. 6c) resulting in relatively shallow thermocline there. The surface waters are also relatively cold in the divergent region south of the equator. The neighborhood of 3N is a relatively warm zone that coincides with the region

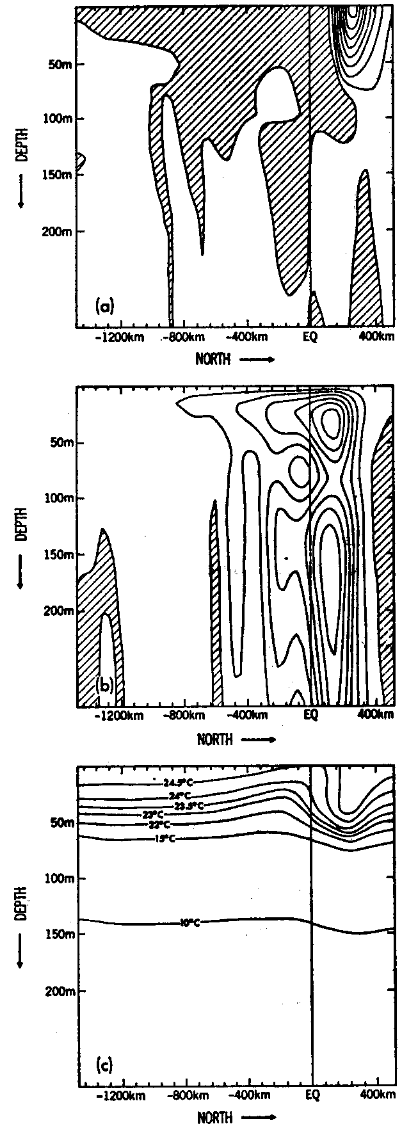


Figure 6. (a) The vertical structure of the zonal flow (isopleths are at 5 cm/sec intervals and regions of westward flow are shaded);

(b) The stream-function for motion in a meridional plane, (contours are at 10^8 m^2/sec intervals and regions of negative values are shaded);

(c) The temperature field in a meridional plane.

All fields are shown 3 weeks after the sudden onset of the winds by which time there is essentially an equilibrium response.

of downwelling. Bakun (1978) has pointed out that the sea surface temperature does indeed have a maximum in the neighborhood of 3N. (See his Fig. 2.) Comparison of meridional sections made in the upwelling and nonupwelling seasons reveals that during the upwelling season the distance from the coast over which isotherms slope upwards to the north in a pronounced manner is about 200 km. (See, for example, the sections in the International Co-operative Investigations of the Tropical Atlantic (ICITA), Equalant I and II, Unesco, Paris, 1973.)

4. Discussion

We have shown that intensification, during the summer, of winds at and south of the equator can intensify the Guinea Current and cause a shoaling of the thermocline near the northern coast of the Gulf of Guinea. Thus are favorable conditions created for upwelling along the coast where the local winds drive an off-shore Ekman drift. This is but one of several mechanisms that contribute to the upwelling. At this stage we can only speculate about some of the others.

Let T be the time it takes for the tropical oceanic circulation to be established, from a state of rest, in response to steady surface winds. (The time T is not a diffusive time scale because once zonal pressure gradients are established in the basin, the winds no longer accelerate the ocean but maintain the pressure gradients. See Philander, 1979). On a time scale long compared to T the oceanic circulation is in equilibrium with the winds. If the winds should fluctuate with a period short compared to T then propagating waves will be excited because the ocean does not have time to come into equilibrium with the rapidly changing winds. If the winds should vary slowly, with a period long compared to T, then the ocean is at any instant in equilibrium with the winds at that instant, and in effect passes through a series of steady states. No waves are excited in this case. (An example of an equilibrium response is a Sverdrup balance in which changes in the flow occur in phase with changes in the curl of the wind-stress.) An equilibrium response is local in time—there is no phase lag between the forcing and the response—but it is not local in space. This is so because the oceanic circulation is closed. Hence a change in one current affects the others. For example, if in the tropical Atlantic only the winds near Brazil have very low frequency fluctuations, then the currents near Brazil *and* those in the Gulf of Guinea will be affected because these various currents form a closed system.

The crucial question is what is the value of the time-scale T? The measurements described by Katz *et al.* (1977) suggest that in the equatorial Atlantic Ocean the value for T is not much longer than the seasonal time-scale. They found that the zonal pressure gradient along the equator in the Atlantic varies practically in phase with the seasonal change in the winds that maintain this pressure gradient.

The upwelling in the Gulf of Guinea—it occurs along the northern coast, the eastern coast, and the equator—is a seasonal phenomenon. In the light of the above discussion we propose that the equatorial oceanic circulation from October until May is such as to favor the presence of a layer of warm surface waters in the eastern part of the basin. Between June and September, when the trade winds intensify steadily, the oceanic circulation changes in such a manner as to cause the disappearance of this warm surface layer. Intensification of the westward South Equatorial Current could contribute to this, for example. In this note we have shown that intensification of the Guinea Current will contribute to the removal of the warm surface waters near the northern coast. At any rate, it appears that to simulate upwelling in the Gulf of Guinea it is necessary to model the seasonal changes in the entire current system in the equatorial Atlantic Ocean.

In an equilibrium response waves are not explicitly excited. The waves however, are always responsible for the adjustment of the ocean from one equilibrium state to another. In a gradual change from one equilibrium state to another waves are not 'seen' because their time-scale is very short compared to that of the change in the forcing function. Suppose that the winds, instead of intensifying gradually between the spring and the autumn, intensify instantaneously on May 1 to their maximum value. For a period of time following May 1, the waves will explicitly be seen to bring about the oceanic adjustment from spring to autumn conditions. The waves are the messengers that carry the message that oceanic conditions are changing. The more gradual the change, the more difficult it is to 'see' the messengers, even though the message remains essentially the same.

Moore *et al.* (1978) have discussed the response of the Atlantic Ocean to the sudden onset of winds near Brazil, and have described the Kelvin wave that propagates into the eastern Atlantic, in detail. They correctly point out that in the wake of this wave, oceanic conditions are more favorable for upwelling. To focus on the Kelvin wave, however, is to focus on the messenger rather than the message. The more gradual the change in wind conditions, the less prominent is the messenger. What is required for a simulation of upwelling in the Gulf of Guinea is a simulation of the change in oceanic conditions between the spring and the autumn (rather than a simulation of the waves that effect this change.)

For a discussion of dynamical similarities between upwelling in the Gulf of Guinea and El Niño events in the Pacific, the reader is referred to Philander (1979).

Acknowledgments. Drs. Orlanski and Ross generously made available to me the numerical program that solves the equations described in Section 2. Mr. Tunison and Ms. Kennedy provided expert technical assistance.

REFERENCES

Arakawa, A. 1966. Computational design for long term numerical integration of the equations of motion: Two dimensional incompressible flow, 1, J. Comput. Phys., *1*, 119–143.

Bakun, A. 1978. The Guinea upwelling. Nature 271, *5641*, 147–150.

Bunker, A. 1977. Monthly surface winds over the North Atlantic. Unpublished report.

Cane, M. A. The response of an equatorial ocean to simple wind-stress patterns. To appear in J. Mar. Res.

Cromwell, T. 1953. Circulation in a meridional plane in the central equatorial Pacific. J. Mar. Res., *12,* 196–213.

Hisard, P. and J. Merle. 1979. Onset of the summer cooling in the Gulf of Guinea. Submitted to Deep Sea Res.

Katz, E. and collaborators. 1977. Zonal pressure gradient along the equatorial Atlantic, J. Mar. Res., *35,* 293–307.

Lilly, D. K. 1965. On the computational stability of numerical solutions of time-dependent, nonlinear geophysical fluid dynamics problems. Mon. Weather Rev. *93,* 11–26.

Moore, D. W. and S. G. H. Philander. 1977. Modeling of the tropical oceanic circulation, *in* The Sea, Vol. VI, E. D. Goldberg, Ed. John Wiley and Sons, Inc., New York.

Moore, D. W., P. Hisard, J. McCreary, J. Merle, J. J. O'Brien, J. Picant, J. M. Verstraete and C. Wunsch. 1978. Equatorial adjustment in the eastern Atlantic. Geoph. Res. Let., *5,* 637–640.

Orlanski, I. and B. B. Ross. 1973. Numerical simulation of the generation and breaking of internal gravity waves. J. Geoph. Res., *78,* 8808–8826.

Orlanski, I., B. B. Ross and L. J. Polinsky. 1974. Diurnal variation of the planetary boundary layer in a mesoscale model. J. Atm. Sci., *31,* 965–989.

Philander, S. G. H. 1979. Variability of the tropical oceans. To appear in Dynamics of Oceans and Atmospheres.

Picaut, J. and J. Verstraete. 1979. Propagation of a 14.7 day shelf wave along the northern coast of the Gulf of Guinea. To appear in J. Phys. Oceanogr.

Received: *14 February, 1978;* revised: *19 September, 1978.*