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Cote: B*6286 Ex: 1

EQUATORIAL ADJUSTMENT IN THE EASTERN ATLANTIC

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Abstract. Observations suggest that the annual upwelling event in the Gulf of Guinea is not associated with changes in the local winds. A possible explanation is that a strong upwelling signal, generated by increased westward wind stress in the western Atlantic, can travel to the eastern Atlantic as an equatorially trapped Kelvin wave. This explanation is analogous to current theories of El Niño in the Pacific Ocean.

Introduction

During the FINE Workshop (FGGE/INDEX/NORPAX Equatorial Workshop, June 27-August 12, 1977) at U.C.S.D., La Jolla, California, the French participants presented data from the equatorial Atlantic and the Gulf of Guinea taken during 1974 (GATE) and 1975. The resulting discussion of these and related data in terms of simple theoretical ideas about equatorial adjustment to time-varying surface winds led to a possible explanation of many of the main features. The primary inputs to the discussion were made by the authors, with occasional comments by other workshop participants.

The data illustrated here are two N-S sections across the equator which were made from the R/V CAPRICORNE at 5°W in January and August, 1975; the time series taken at 0°N, 10°W by Rybnikov from the USSR vessel PASSAT (including moored current meter data subsequently analyzed by Duing); 1974 sea level and SST data at various coastal locations along the Gulf of Guinea (Picaut and Verstraete, personal communication); and the mean sea surface slopes along the equator based on historical data (Neumann, et al., 1975). Basic ideas about baroclinic equatorial adjustment in the ocean, in the presence of continental boundaries, are contained in Moore (1968); Lighthill (1969); O'Brien and Hurlburt (1974); Hurlburt et al. (1976); McCreary (1976); Moore and Philander (1977); and Cane and Sarachik (1976, 1977).

Observations

Figure 1 (a,b) shows the temperature and zonal velocity data obtained by Hisard at 5°W in January, 1975. Note the strong eastward undercurrent with maximum velocity in excess of 120 cm/sec at 70 meters depth. There is a marked depression of the 15°-20°C isotherms in

the vicinity of the undercurrent core. The zonal velocity at the surface near the equator is westward and relatively weak (10-20 cm/sec). Figure 2 (a,b) shows the same data for August, 1975. The undercurrent is weaker and shallower, with a maximum velocity of ~80 cm/sec at 40 meters depth. The 16°-21°C isotherms now show a marked upward bowing in the vicinity of the undercurrent. The westward surface flow at the equator is much stronger (60-100 cm/sec). It is clear from comparing the January and August data that a significant change has occurred in the equatorial temperature and flow fields at 5°W during the intervening period.

Figure 3 (a) shows the time series of SST observed at 0°N, 10°W by the PASSAT during 27 June-16 July, 27 July-15 August, and 29 August-19 September, 1974. Figure 3 (b) is the time series of zonal velocity as a function of depth from moored current meters at the same location, for the entire period, 24 June-13 September. There is a marked drop (~2°C) in SST between 6 July and 9 July. There is also an upward shift in the core of the undercurrent, starting perhaps as early as 28 June. We hypothesize that the upward motion of the undercurrent and the appearance of cold water at the sea surface are associated with an upwelling event. The vertical migration of the undercurrent core begins before the

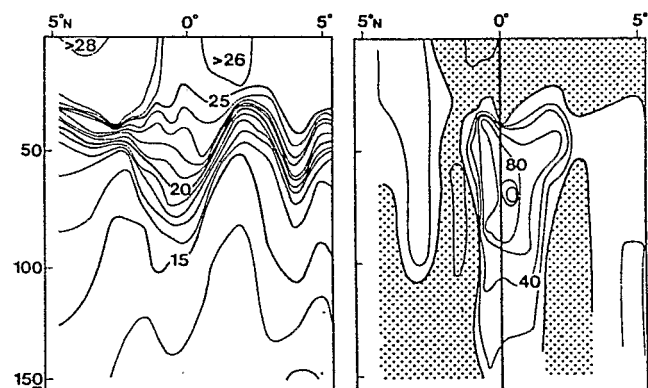


Fig. 1 (a). Temperature (°C) as a function of latitude and depth at 5°W in January, 1975.

Fig. 1 (b). Zonal velocity (cm/sec⁻¹) as a function of latitude and depth at 5°W in January, 1975. The unmarked contour at the core of the Undercurrent is 120 cm/sec⁻¹. Currents are measured relative to 300 meters.



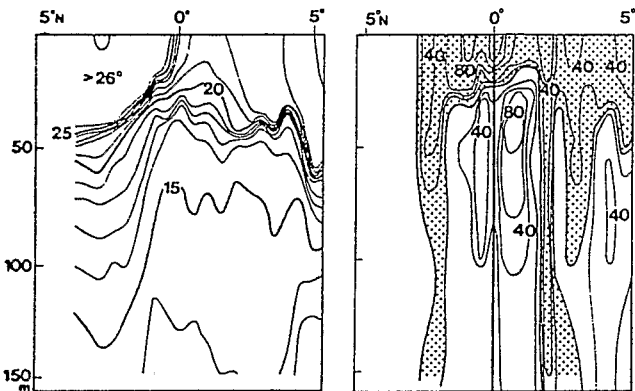


Fig. 2 (a). Temperature ($^{\circ}\text{C}$) as a function of latitude and depth at 5°W in August, 1975.

Fig. 2 (b). Zonal velocity (cm/sec^{-1}) as a function of latitude and depth at 5°W in August, 1975. The unmarked contours in the westward surface flow at the equator are 100 cm/sec^{-1} . Currents are measured relative to 300 meters.

marked change in SST, since the mixed layer must be advected away before the cold water can appear at the surface. It is clear from the PASSAT data that the upwelling event as revealed by the cold SST was of at least two months' duration.

Figure 4 (Picaut and Verstraete, personal communication; Houghton and Beer, 1976) shows time series of daily sea surface temperature at a number of coastal stations in the Gulf of Guinea. The temperature at all locations begins a gradual decline which seems to start about 1 June; an extended Tema record, provided by Hisard, shows that this decline actually began about 20 May. There is a marked drop ($\Delta T \sim 6^{\circ}$ at Tema) in SST beginning about 13 July. According to Houghton (1976) isopycnals remained relatively flat prior to 12 July to at least a depth of 100 meters. Between 12 July and 18 July, the isopycnals rose about 50 meters, the most rapid increase taking place after 15 July. A shallow thermocline at 40 meters depth was completely brought to the surface by 20 July. (The surfacing of the thermocline is evident in Figure 4 by the appearance of a 15-day baroclinic tidal oscillation. This signal occurs in corresponding sea level records throughout the year, but can exist in the SST records only when the thermocline is near the surface.) An upwelling event similar to the equatorial one has occurred along the African coast.

There is considerable evidence that upwelling events, similar to those illustrated in Figure 1-4, occur every year. Bakun (1978) has analyzed historical SST records in the Gulf of Guinea. One of his figures shows that in August a narrow band of cool ocean temperature exists all along the equator, as well as between 5°E and 10°W along the African coast. Merle (1977) has investigated seasonal variations of the deeper thermal structure in the tropical Atlantic using hydrographic data on file at NODC. According to Merle, in the eastern Atlantic cooling gradually begins in

May and reaches an extreme in August or September. The cooling is identifiable to depths of 150-200 meters in some cases, but is generally confined to the upper 100 meters; it is clearly associated with an uplifting of the thermal structure of the order of 25-50 meters. Using the value of salinity at the core of the Undercurrent as a measure of the flow, he argues that from 0° - 20°W the speed of the Undercurrent reaches a minimum in September, a principal maximum in April-May, and a strong secondary maximum in December-January.

The most interesting aspect of the upwelling in the Gulf of Guinea is that it is not related in an obvious way to the local winds (Houghton, 1976; Bakun, 1978). According to Hastenrath and Lamb (1977), the winds in the equatorial belt east of 10°W are primarily meridional with little variation in the zonal wind from month to month. The standard explanation of upwelling requires changes in zonal (alongshore) wind at the equator (coast), and subsequent Ekman divergence, to induce upwelling. It is not likely, then, that the marked differences in the thermal and velocity structure of the ocean discussed above are caused by changes in the local wind.

Katz et al. (1977) suggest that the upwelling signal at 10°W in the summer of 1974 (Figure 3) was dynamically related to events which took place far to the west. They report that the pressure gradient between 10°W and 35°W increased markedly between mid-June and the end of August. In fact, this increase of the zonal pressure gradient is part of an annual cycle. Figure 5, from Neumann et al. (1975), shows mean slopes of dynamic height along the equator based on historical data. It is evident in the figure that there is a substantial increase in the zonal pressure gradient between the February-March and July-September-November sections. Merle (1977) corroborates this result and finds that the pressure gradient is weakest in April (5 dynamic cm) and largest in September (30 dynamic cm).

The fluctuation of equatorial pressure gradient occurs in conjunction with large changes in the zonal wind field west of 10°W . Katz et al. (1977) show the monthly averaged

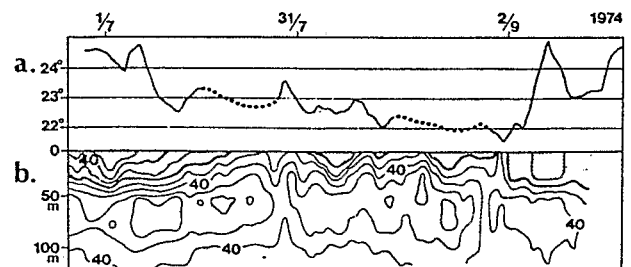


Fig. 3 (a). Time series of SST on the equator at 10°W during July-September, 1974.

Fig. 3 (b). Time-versus-depth contours of zonal velocity at the equator from moorings at 10°W during July-September, 1974. The unmarked contours at the core of the Undercurrent are 80 cm/sec^{-1} .

wind stress on the equator between 10°W and 40°W, as computed by Bunker. The westward wind stress is at its minimum value in March and increases rapidly from .25 dyne/cm² to .55 dyne/cm² during April, May and June. It remains at this high level until November. As the westward wind increases, the pressure gradient quickly follows reaching a maximum several months after that of the wind.

A Simple Explanation

We hypothesize that the above oceanic observations represent the response of the equatorial Atlantic Ocean to the strengthening of the westward zonal wind stress in the western Atlantic. Teleconnections between the eastern and western Atlantic can be simulated with simple linear models. Suppose that a wind stress confined to the western ocean is switched on at some initial time; then the models predict that:

- 1) In the region where the (westward) zonal wind intensifies, surface water piles up against the western boundary, and as a result a zonal pressure gradient is established which balances the applied wind stress (see Figure 5).
- 2) To the east of the forcing, a Kelvin wave upwelling is generated, propagating eastward at a speed $c = \sqrt{g'H}$, of the order of 250 cm/sec (Moore and Philander, 1976), where g' is reduced gravity and H is the depth of the

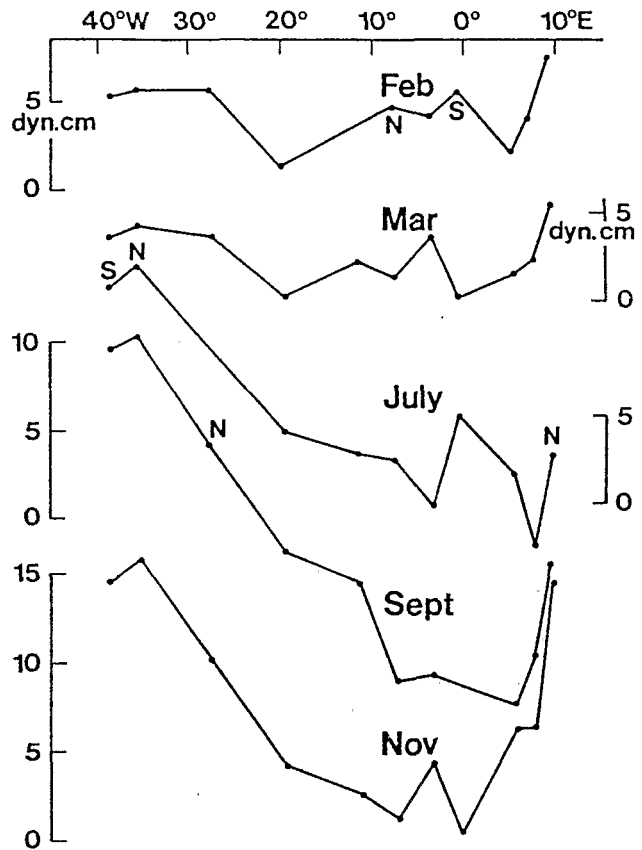


Fig. 5. Mean sea-surface slopes along the equator based on historical data.

- 3) When this upwelling disturbance reaches the eastern boundary, it splits with a signal going poleward along the coast in both hemispheres (see Figure 4).

An accompanying note by O'Brien, et al. (1978) illustrates these ideas using a model in an ocean basin which includes the African coast. This adjustment process is completely analogous to a currently proposed explanation for the El Niño phenomenon in the Pacific (Hurlburt, et al., 1976; McCreary, 1976).

Discussion

The transient model response discussed in the previous section suggests that a definite time lag must exist between the onset of equatorial and coastal upwelling: Suppose that an equatorially trapped event is originally observed at 10°W. It must travel 19° of longitude to reach the African coast. After it reflects from that boundary, in part as a coastally trapped Kelvin wave, it must travel back another 10° to reach Tema. Therefore,

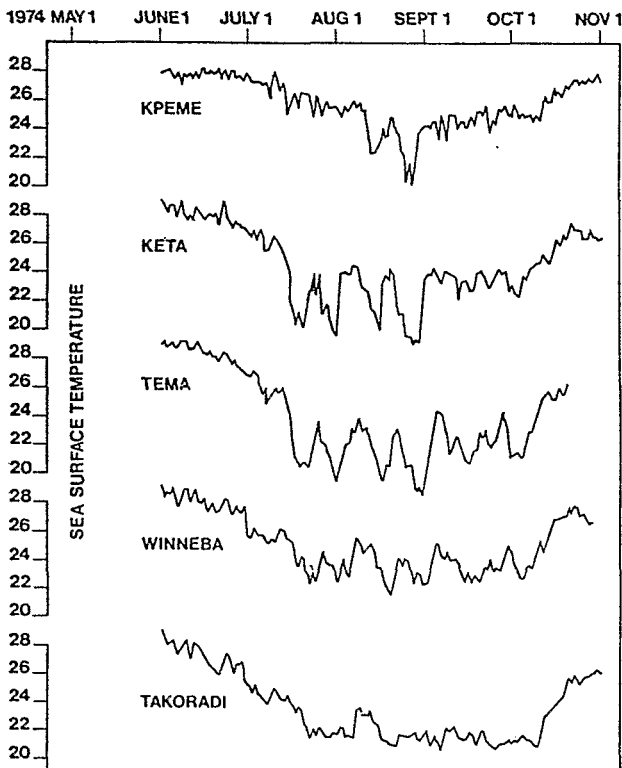


Fig. 4. Time series of SST (°C) from coastal stations in the Gulf of Guinea during 1974. The data from Ghana were kindly provided by the Ghana Fisheries Research Unit in Tema.

the signal at Tema should lag the equatorial one by roughly 12 days. Such a lag may have existed during the 1974 upwelling event; however, it is difficult to determine a moment on the onset of upwelling in either region. Did the equatorial upwelling begin on 6 July with the rapid dropping of SST, or on 28 June with the rise of the core of the Undercurrent? Did the coastal upwelling begin on 12-15 July with the rapid rise of isopycnals, or on 1 June with the gradual cooling of SST? More observations are needed.

Hisard (personal communication) has pointed out that in 1974 a gradual lowering of SST at Pointe-Noire (4°49' south of the equator) began about 15 May, considerably before the rise of the Undercurrent at 10°W and slightly preceding the gradual drop of temperature at Tema. McCreary and Moore (personal communication) have recently been studying the response of simple models to winds oscillating at frequency ω . They find that the obvious phase relationships connecting events in the transient response are altered considerably by the reflection of Rossby waves from a meridional boundary. It may be possible to account for the early upwelling in this way.

If fluctuations of local wind are not the cause of upwelling in the Gulf of Guinea, what then is? The point of this note, and its companion, is to suggest a possible answer to this puzzle. Although the available data as yet do not allow a thorough test of our hypothesis, they do suggest the possibility that teleconnections across the Atlantic are occurring seasonally in much the same way that they are known to occur in the Pacific on interannual (El Niño) time scales.

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(Received March 20, 1978;
revised May 22, 1978;
accepted June 14, 1978.)