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THE OCEANIC CIRCULATION OF THE TROPICAL ATLANTIC, AND ITS VARIABILITY, AS OBSERVED DURING GATE

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Abstract--The principal components of the surface current system in the tropical Atlantic are the westward South Equatorial Current south of 3°N, the eastward Equatorial Countercurrent between 3°N and 10°N and the westward North Equatorial Current north of 10°N. The subsurface, eastward Equatorial Undercurrent is confined to about 1 $1/2^{\circ}$ latitude and is centered on the equator. These mean currents are subject to fluctuations over a spectrum of frequencies. Very high-frequency, turbulent fluctuations are of major importance in the mixed layers at the ocean surface and below the core of the Undercurrent. One-dimensional models cannot simulate these mixed layers because they are strongly influenced by the divergence of large-scale currents. Fluctuations with periods less than the inertial period correspond to inertia-gravity waves but their spectral properties, near 8° N, are unusual in two respects:

(i) at periods between 1/2 hr and 10 min there are very energetic oscillations associated with a thermocline-trapped internal mode;

(ii) amplitudes of inertial waves below the thermocline are correlated with the intensity of surface winds. Near 5°N the inertial peak in the spectrum disappears and equator ward of 5° latitude equatorially trapped waves dominate the spectrum at periods longer than 10 days. There are hints that wave-like fluctuations do not have a universal spectrum in low latitudes. Particularly

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energetic equatorial oscillations observed during GATE included the following: 3- to 5-day equatorially trapped inertia-gravity waves which were forced by atmospheric disturbances with the same period; 16-day meanders of the Equatorial Undercurrent which may be related to atmospheric fluctuations with the same period; 30-day, 1000 km waves which appear to be due to instabilities of the surface currents. Superimposed on these oscillations is a trend that is part of the seasonal cycle: for example, the zonal pressure gradient in the equatorial plane increased throughout GATE, practically in phase with the intensification of the tradewinds. This information sheds light on the seasonal upwelling in the Gulf of Guinea which is not correlated with changes in the local winds.

1. INTRODUCTION

The field phase of GATE took place during the boreal summer of 1974. By that time the principal features of the oceanic circulation in the tropical Atlantic had been established, but essentially nothing was known about the variability of this circulation. It was the objective of the oceanographic component of GATE to determine the temporal and spatial scales of variability in the tropical oceans, and to relate the observed oceanic variability to that of the atmosphere. This paper attempts to summmarize the main results from the oceanographic experiment. It should be kept in mind that the analysis of the GATE data is, at this time, far from complete. We shall try to identify further problems that might be investigated in the future. We unfortunately are unable to reference all the interesting GATE-related studies. (The GATE oceanographic atlas, Düing et al., 1980, which is in preparation, will include a complete list of publications.)

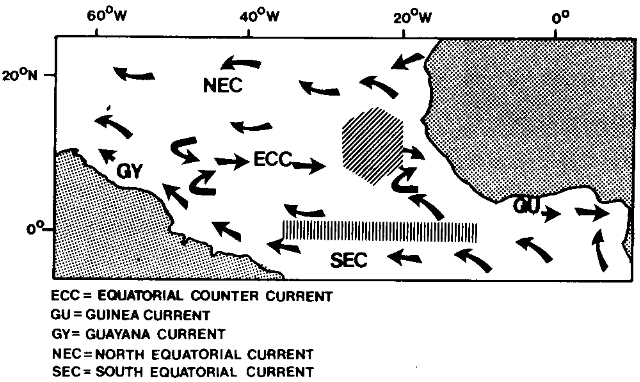

Fig. 1. Shaded areas indicate regions of intensive oceanographic measurements during GATE. Surface currents shown schematically.

Figure 1 shows that during GATE regions of intensive measurements were clustered about the equator, and around location 8°N, $23^{\circ}30'$W; measurements along the meridian $23^{\circ}30'$W linked these two areas. For logistical reasons, the observational program was most intense during three three-week periods: Phase I from 26 June until 15 July; Phase II from 28 July until 17 August; and Phase III from 30 August until 19 September. Philander et al. (1974) describe the oceanographic program in detail; for summaries of these technical reports see Düing (1974) and Philander (1974).

Given the relatively short duration of GATE - the longest available time-series covers 86 days - it is possible to describe variability up to a period of about 1 month only. However, the seasonal cycle in the tropical Atlantic is sufficiently strong for it to appear as a trend in the GATE data set. Viewed in the context of historical data, this trend provides some information about the seasonal cycle. Figure 2 is a useful frame within which to discuss variability over this range of frequencies. It shows, as a function of latitude, the periods at which oceanic motions are likely to be wave-like and the periods at which motions are likely to be in equilibrium with atmospheric forcing. In summarizing the GATE results we shall try to assess the extent to which the observations corroborate the theoretical ideas that lead to the various domains in Figure 2, which is based on theoretical considerations by Philander (1979a).

This paper is divided into six parts: Section 2 describes mean oceanographic conditions during GATE itself, as well as the GATE observations in the context of the seasonal cycle; Section 3 concerns upwelling in the Gulf of Guinea; Section 4 deals with wave-like motions observed during GATE; mixing processes in the surface layers and below the core of the Equatorial Undercurrent form the subject of Section 5; Section 6 is a summary and discussion of the results.

2. THE MEAN OCEANOGRAPHIC CONDITIONS AND THEIR SEASONAL VARIABILITY

A. Mean Conditions

The oceanic circulation is maintained primarily by the wind-stress on the ocean surface. Elsewhere in this volume Hellerman (1979) describes the seasonal variability of the winds over the tropical Atlantic, and Krishnamurti and Krishnamurti (1979) describe their variability during GATE. The winds are primarily from the south over the Gulf of Guinea (east of 10°W). Trade winds prevail west of 10°W: northeast trades north of the intertropical convergence zone (ITCZ); southeast trades south of the ITCZ. Of particular interest to us is the steady intensification, between June and September, of the winds west of about 30°W. This intensification is associated with a northward migration of the ITCZ.

Figure 1 shows the surface currents schematically: the eastward North Equatorial Countercurrent, between 3°N and 10°N approximately, is sandwiched between the westward North Equatorial Current to the north and westward South Equatorial Current to the south. The Countercurrent is fed by the Brazilian Coastal Current at its western

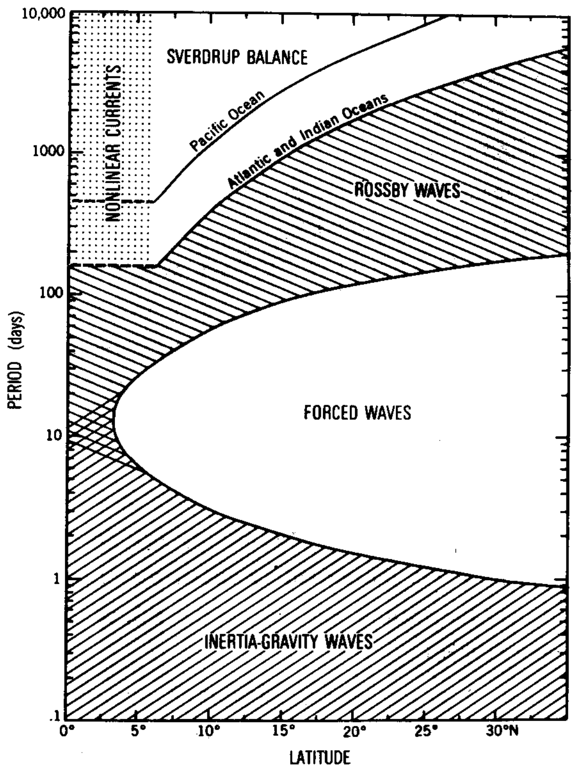

Fig. 2. The nature of the oceanic response, as a function of latitude and period, for the first baroclinic mode (after Philander, 1979a).

extreme (Vazquez de la Cerda, 1978), has a transport of approximately $17 \times 10^6\ m^3\ s^{-1}$ across 40^oW (Cochrane, 1978), and weakens further to the east. Its vertical structure across 23.5^oW is shown in Figures 3, which depict mean oceanographic fields observed during GATE. It can be shown that the Countercurrent is nearly in geostrophic balance and it is associated with the strong latitudinal slope of the isopycnals between 3^oN and 10^oN; computed geostrophic velocities and directly measured currents have been shown to compare well (Bubnov et al. , 1979). In the mixed surface layer there is, however, a significant non-geostrophic component which is presumably associated with the Ekman layer (Halpern, 1979).

An intriguing feature of the hydrographic data shown in Figures 3 is the steep slopes of the isopycnals near 5^oS (at depths between 200 m and 500 m) and near 6^oN (at depths between 200 m and 400 m). These slopes imply subsurface eastward currents documented by Hisard et al. (1976).

Symmetrically about the equator flows the subsurface, eastward Equatorial Undercurrent which had an average maximum speed of the order of 75 cm s^{-1}. Transport estimates of the Undercurrent must be interpreted with some caution because of the large variability encountered over short time periods. All estimates fall within the range $(12.5 \pm 2.5) \times 10^6\ m^3\ s^{-1}$ for transport within the 20 cm s^{-1} isotach.

Based on five sections, occupied once each, Bruce and Katz (1976) found the transport to increase downstream as far as 16°W, but further eastward, where the zonal pressure force along the equator changes from eastward to westward, the transport rapidly decreased. Associated with the Undercurrent is a core of high salinity water. (In Figure 3c, it appears somewhat to the south of the equator.) The salinity maximum is, on the average, about 20 m above the core of the zonal velocity. This is probably a consequence of the strong equatorial upwelling. Eastward of 28°W there is a gradual salt flux out of the Undercurrent. This loss of salt is due to upwelling into the divergent surface layers, and downward entrainment into the pycnostad. From the absence of downstream, latitudinal spreading of the salinity core, Katz et al. (1979) infer a meridional convergence of fluid into the Undercurrent. This meridional circulation is the subject of theories for the Undercurrent and is thought to be important in its momentum balance (Moore and Philander, 1977).

Below the Undercurrent down to a depth of 1000 m, the flow was highly variable and the zonal velocity component reversed direction during GATE. Philander's (1973) steady state model for the equatorial currents requires westward flow at depth and is inconsistent with these observations. It is conceivable that the variability of the deep currents plays a central role in determining the mean conditions.

The various currents described thus far are associated with density gradients such as those in Figure 3d. The thermocline slopes longitudinally and increases in depth from east to west along a circle of latitude (Fig. 4). Sea surface temperatures are highest where the thermocline depth is greatest (close to the South American coast as shown in Figure 5a). The variability of the sea surface temperature is largest where the thermocline is shallow: near the equator and in the eastern part of the Gulf of Guinea (Fig. 5b).

B. Seasonal Variability

In the introduction we mentioned that the seasonal cycle in the tropical Atlantic is sufficiently strong to appear as a trend in the GATE data. Studies of this low frequency cycle have motivated investigators to re-examine the historical data set so that a fair amount is now known about the seasonal cycle.

Between June/July and August 1974 the zonal pressure gradient in the upper ocean, along the equator in the western and central Atlantic, nearly doubled. This seems to be

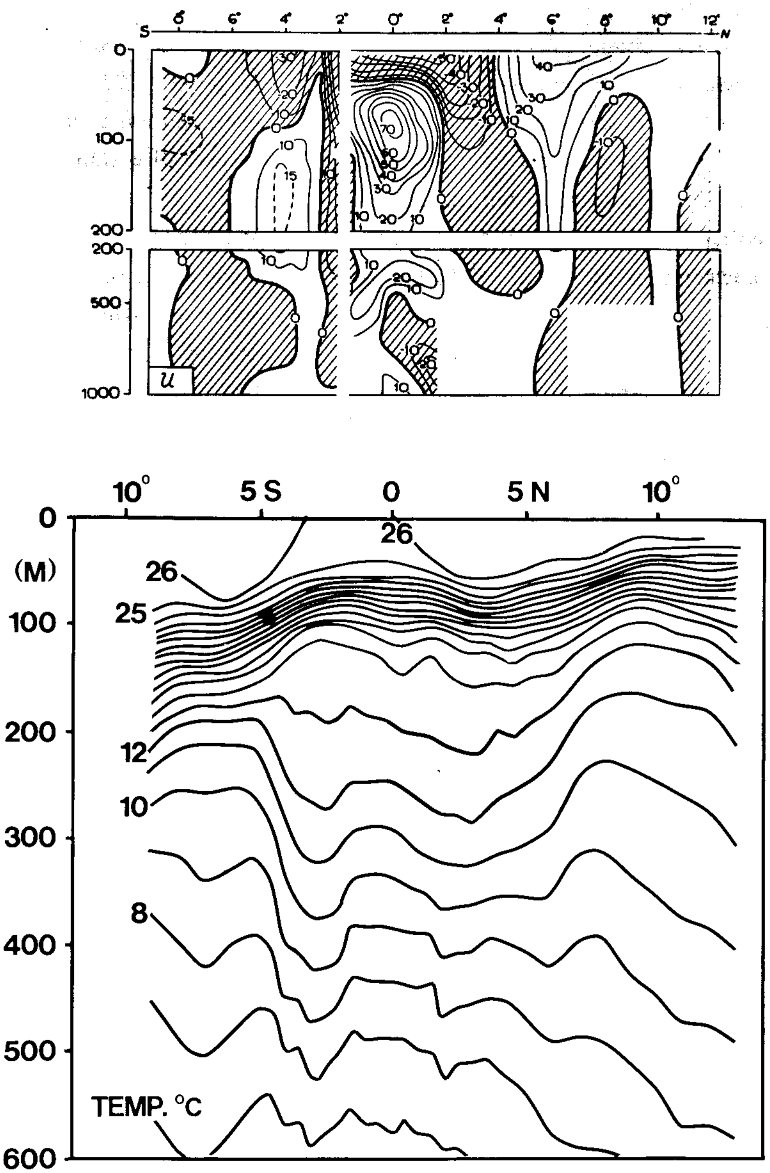

Fig. 3. a. The mean zonal currents across 23.5°W on the basis of direct measurements during GATE north of 2°S and on the basis of geostrophic calculations south of 2°S (after Bubnov et al., 1979). b,c,d. The mean fields of temperature, salinity and σ_t, respectively, on the basis of measurements from R/V TRIDENT and R/V SEMEN DEZHNEV.

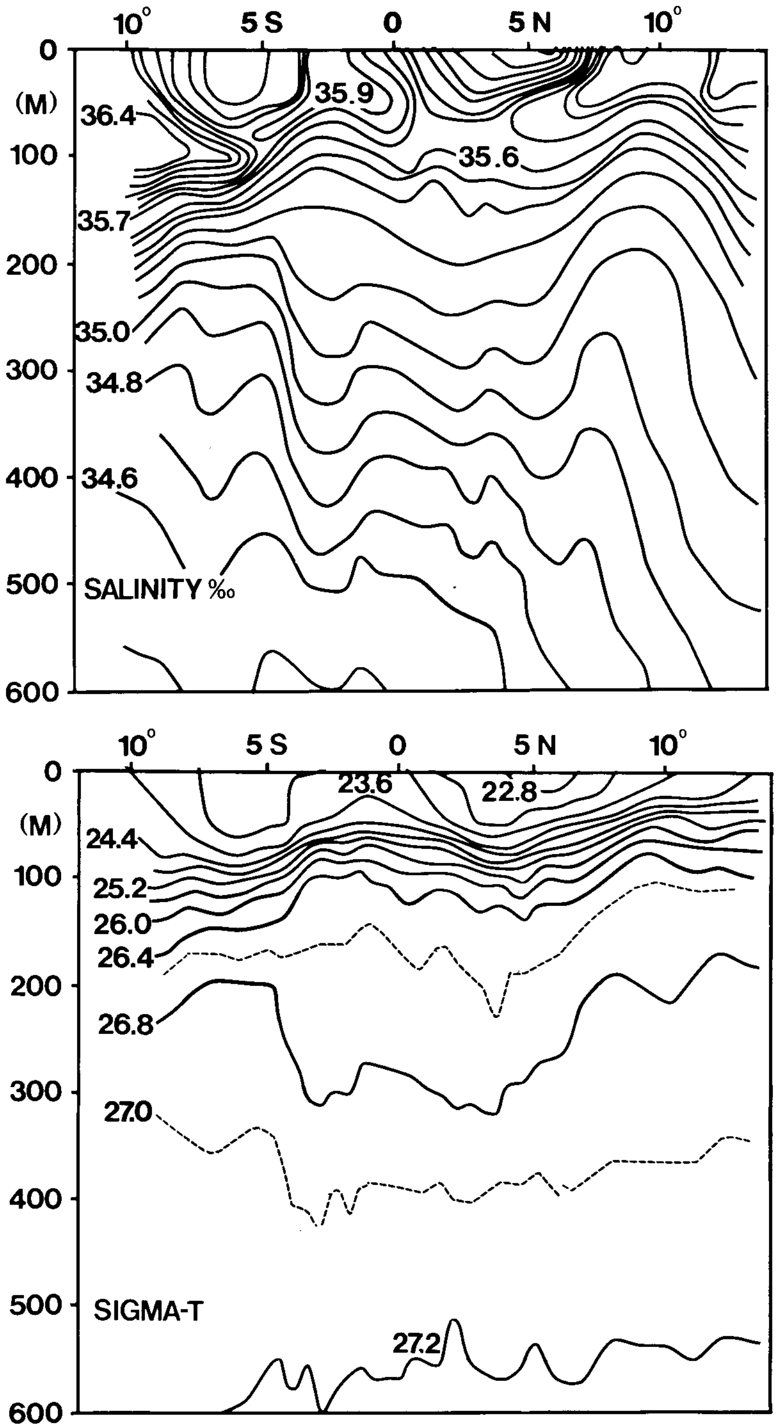

Fig. 3. (continued)

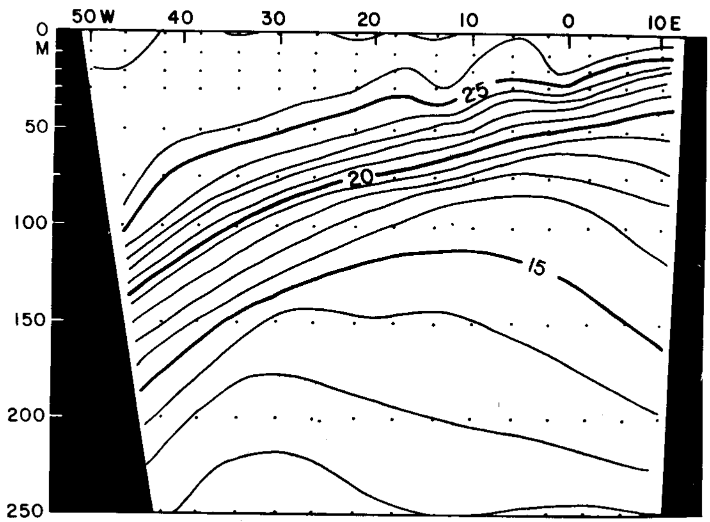

Fig. 4. Temperature section along the equator.

part of a seasonal trend from low values in the boreal spring to high values in mid-summer and fall. This trend is paralleled by the zonal wind-stress over the western equatorial Atlantic (Katz et al., 1977; Neumann et al., 1975; Merle, 1978). This result is remarkable because it suggests that on the seasonal time-scale, the response of the equatorial Atlantic is in equilibrium with the forcing function. Apparently the regime boundary for the equilibrium response of the Atlantic Ocean in Figure 2 is quite realistic near the equator. Next, one would like to know how the equatorial currents changed in response to the changing winds and zonal pressure gradient. Surprisingly, the measurements do not show a distinct trend with increasingly larger velocities and transports during GATE. The westward surface flow at the equator was about the same in Phases I and II and decreased somewhat during Phase III. The Undercurrent transport during Phase II was apparently smaller than during Phases I and III (Bubnov and Egorikhin, 1977). It is extremely difficult to quantify these statements. There are several reasons for this. Superimposed on the low frequency (seasonal) trends in the data are energetic fluctuations with periods of a month

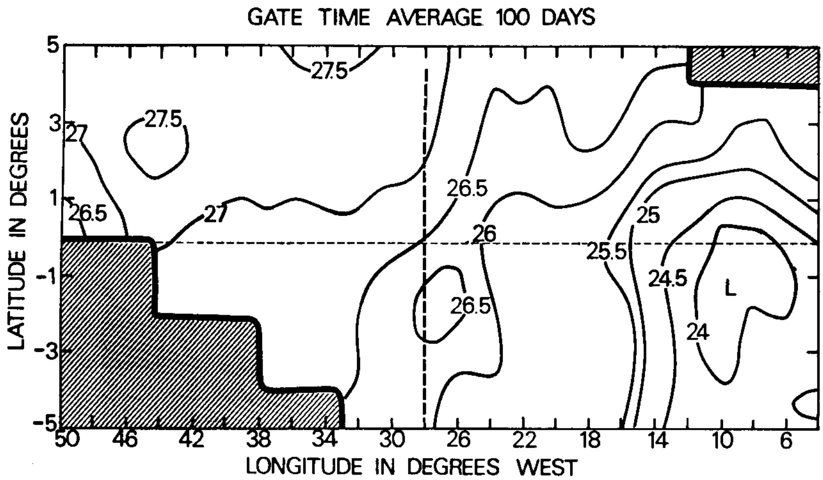

Fig. 5a. The mean of sea-surface temperature during GATE (after Brown, 1979).

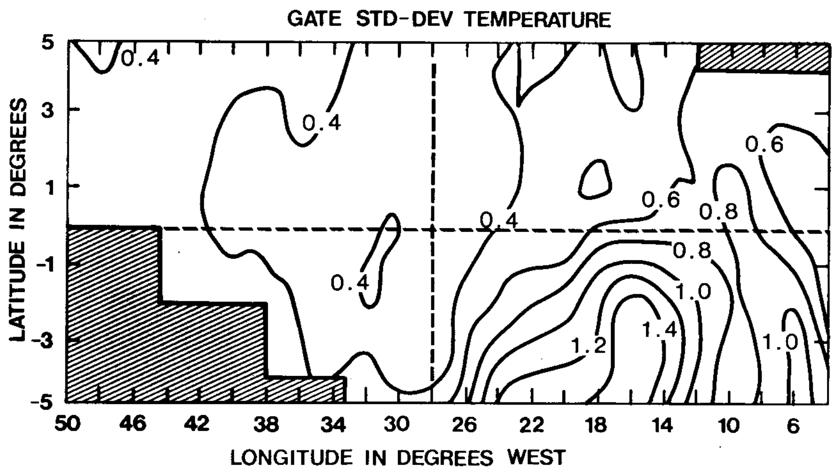

Fig. 5b. The standard deviation of sea-surface temperature during GATE (after Brown, 1979).

and less. The narrow Equatorial Undercurrent is difficult to resolve with a few moorings, hence it is not a simple matter to estimate variations in transport. Finally, the longest time-series are from surface moorings which are sensitive to the sea-state, and which tend to overestimate velocities (Gould and Sambuco, 1975). According to Bubnov and Egorikhin (1977) the Undercurrent transport across 23.5°W was 16.7 Sv in Phase I. Though care must be taken not to attach too much significance to the difference between these various numbers, it is intriguing that a lower Undercurrent transport in Phase II, than in Phases I and III, is consistent with the semi-annual oscillation of the salinity core in the western equatorial Atlantic which Neumann et al.(1975) and Merle (1978) find in the historical data. We are assuming (see also Hisard and Merle, 1979) that there is a correlation between the salinity and the intensity of the Undercurrent (which transports saline water eastward). But this assumption may be questionable. Katz et al. (1979) noted that the salinity is determined not only by the intensity of the Undercurrent, but also by the intensity of upwelling into the divergent surface layers. Only further measurements can clarify the seasonal cycle of oceanographic parameters near the equator.

In summary, it appears that in response to the uniform increase in the intensity of the westward winds over the western equatorial region during GATE, the zonal pressure force along the equator increased steadily. Neither the westward surface flow at the equator, nor the eastward Equatorial Undercurrent, changed in a correspondingly uniform and marked manner. From the correlation between the wind-stress and the zonal pressure force one could infer that, on a seasonal time-scale, the ocean is in equilibrium with the changing winds. This would imply that the ocean in effect passes through a series of steady states, each of which is in equilibrium with the wind at that instant. From Charney's (1960) calculations we know that the more intense the wind, the more intense is the (steady-state) eastward Undercurrent, which becomes a surface current for strong winds. Such a model gives a hint as to why the westward surface flow did not intensify during GATE, but the model also predicts a uniform intensification of the Undercurrent which is inconsistent with the GATE data. It therefore seems that steady-state theory does not apply. In other words, the equilibrium time-scale T (the upper curves in Figure 2) is not short compared to the seasonal time-scale. We conclude from the GATE data that T is probably comparable to the seasonal time-scale.

From Figure 6 it is evident that the geostrophic component of the Equatorial Countercurrent between 3°N and 10°N changed significantly during GATE. Were these changes in phase with changes in the surface winds, thus implying an equilibrium response? (Figure 2 suggests that on a seasonal time-scale an equilibrium response is possible in the Atlantic Ocean.) It will also be interesting to know whether the transport of the Countercurrent is in accordance with Sverdrup's theory. Further analyses of GATE data along these lines will be of great value.

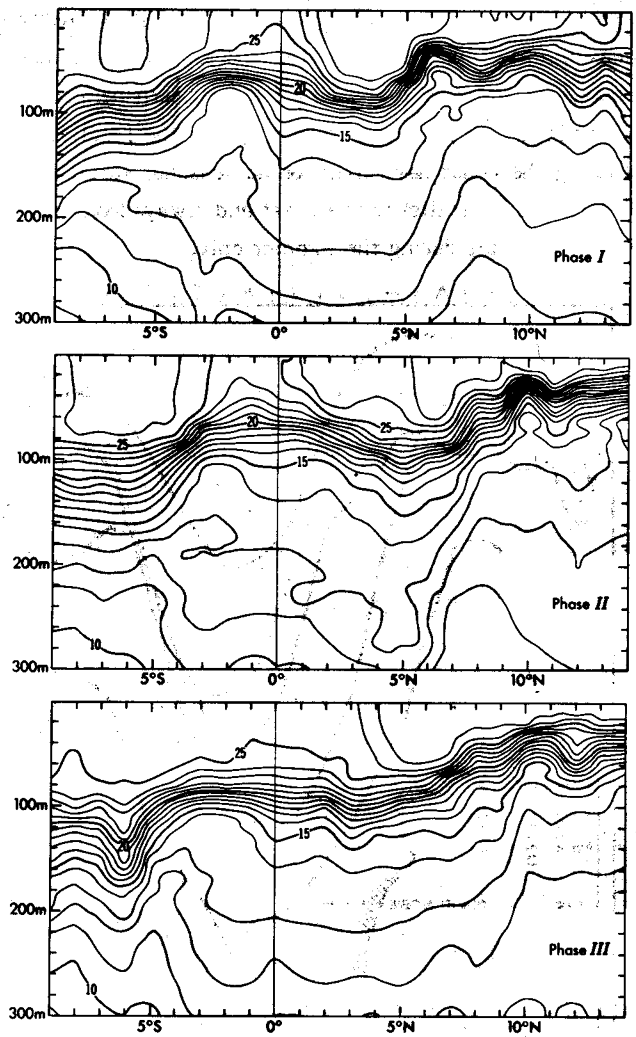

Fig. 6. The mean temperature along 23.5°W for each Phase of GATE (courtesy J. Merle).

3. UPWELLING IN THE GULF OF GUINEA

Though not a region of intensive measurements during GATE, the Gulf of Guinea has recently attracted considerable attention because of the intriguing upwelling phenomenon along the northern and esastern coast. Figure 7 shows the monthly mean sea surface temperature as measured at a number of coastal stations. The sea surface temperature is seen to have a pronounced seasonal low during the (northern hemisphere) summer. Oddly enough, this upwelling is not correlated with changes in the local surface winds. In fact, the surface winds along the coast have little or no seasonal variability (Verstraete et al., 1979); these winds are always parallel to the coast and always favorable for upwelling. Yet upwelling-related cooling occurs during the summer only.

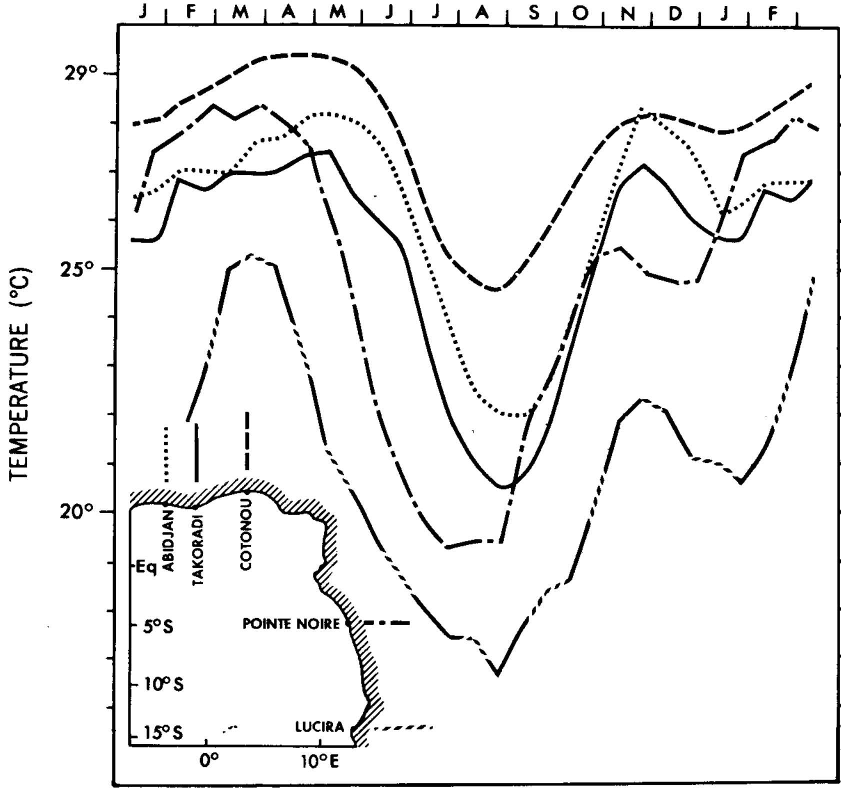

Fig. 7. The mean monthly sea surface temperature at coastal stations along the Gulf of Guinea (after Hisard and Merle, 1979).

Hisard and Merle's (1979) analysis of historical data, and data obtained during GATE, shows that the upwelling comes to an end in September when a layer of warm fresh water covers the entire eastern tropical Atlantic. As a consequence, the thermocline is at such a depth that the local winds cannot induce low sea surface temperatures. In May, this warm water disappears, cold saline water is exposed, the thermocline is relatively shallow, and equatorial and coastal upwelling is possible. What causes the appearance and disappearance of the layer of warm surface waters?

Katz et al. (1977) found that, on a seasonal time-scale, the slope of the thermocline in the equatorial plane is subject to enormous changes. This implies that there is a considerable seasonal redistribution of heat, on a basin-wide scale, in the equatorial Atlantic Ocean. This suggests that seasonal changes in advective patterns could be responsible for the Gulf of Guinea phenomena described above. Ultimately the surface winds cause these changes and the winds over the entire basin (and not the local winds only) come into play. To understand upwelling in the Gulf of Guinea, one would have to understand the seasonal cycle in the entire equatorial Atlantic Ocean. Philander (1979b) has studied part of this problem by showing that intensification of the eastward Guinea Current (during the summer) will lead to a shoaling of the thermocline near the northern coast of the Gulf of Guinea and hence to conditions that favor upwelling.

According to the arguments given above, oceanic conditions during the non-upwelling season (state A say) are a consequence of the circulation associated with the winds that prevail during state A. Upwelling conditions (state B say), are caused by the (non-local) winds that prevail during state B. Waves effect the adjustment from state A to state B. If the change from A to B is abrupt then the waves are prominent, as in the models of Moore et al. (1978), O'Brien et al. (1978) and Adamec and O'Brien (1978). If the change is gradual, the waves may not be evident but are nonetheless responsible for the adjustment. The waves, however, are not responsible for either state A or B but only for the adjustment from the one to the other. To say that waves cause the upwelling is equivalent to holding a messenger responsible for his message. There exist several studies of the adjustment of the equatorial Atlantic Ocean from one state to another but these are not studies of the upwelling phenomenon as such.

4. WAVES

Figure 2 indicates that the response of the extra-equatorial oceans is in the form of inertia-gravity waves at high frequencies and is in the form of Rossby waves at lower frequencies. At intermediate frequencies no freely propagating waves are possible in non-equatorial latitudes. Thus far the analyses of the GATE time-series from non-equatorial latitudes have focused on inertia-gravity waves. Further analysis could provide information about the forced wave regime but the time-series are too short for a study of extra-equatorial Rossby waves. The analysis of data from equatorial latitudes has so far concerned periods longer than a few days. A study of higher-frequency fluctuations there should prove valuable. In this section we first describe some of the properties of the spectrum of equatorial and non-equatorial waves; thereafter we discuss two specific frequencies: the energetic 16-day and one-month oscillations observed near the equator.

A. Extra-Equatorial Inertia Gravity Waves

Between the inertial frequency (f) and the Brunt-Vaisala frequency (N), the kinematics of oceanic motion correspond to a linear superposition of inertia-gravity waves (Müller et al., 1978), but the energetics of the flow is strongly non-linear. In non-tropical oceans the non-linear interactions redistribute energy across the frequency band f to N so efficiently that the spectral properties of the flow are always the same, and are described by the Garrett and Munk (1975) spectrum. Because of the non-linear interactions, it is difficult to locate a source region for inertial-gravity waves; within a few wavelengths from a generation area the universal shape of the spectrum can be restored (Wunsch, 1976). Of particular interest to us is the absence of any correlation between the energy levels of inertia-gravity waves and the energy levels of the atmospheric motion at corresponding frequencies.

As the equator is approached, the inertial frequency tends to zero. One can speculate that as the freqency band f to N widens, it is less likely to be saturated, so that there may be departures from the Garrett-Munk spectrum. The question then is whether the GATE data shed light on this conjecture.

An analysis of current meter records from 8°49.9'N, 22°52.6'W reveals unusually high energy levels at periods between 10 min and 1/2 hr (Käse, 1979). This frequency range is associated with the gravest internal mode which is trapped in the thermocline. The high-frequency fluctuations that Proni, Ostapoff and Sellers (1978) observed to be coherent over the upper 100 m presumably correspond to the same mode. Käse argues that the amplitude of this mode is limited by wave breaking. This energetic mode which causes the energy spectrum to differ from the Garrett-Munk spectrum may be common to regions with thermoclines as shallow and strong as the one in the neighborhood of 8°N 22°W. The concept of a 'universal' spectrum may therefore still be valid, but it will be different for different geographical areas. As one moves southward across the North Equatorial Countercurrent, from 8°N to 3°N say, the depth of the thermocline increases significantly (see Fig. 3). It will be interesting to know how this affects high-frequency internal waves. Further analysis of GATE data and future theoretical work could answer this question.

Käse and Olbers (1979) have studied waves with frequencies slightly higher than the inertial frequency, at 8°N 22°W, in the upper ocean at depths across the thermocline. They find that fluctuations in the amplitude of these waves just below the thermocline correlate with fluctuations in the wind-stress amplitude at corresponding frequencies. This result is surprising in the light of our earlier comments about how difficult it is to locate a generation region for inertia-gravity waves in non-tropical latitudes. Because of the very strong thermocline at 8°N, the downward propagating waves observed below this thermocline would have originated a considerable horizontal distance away, in the surface layers. The atmospheric fluctuations at that distant point will correlate with those at the point of observation because the surface winds are particularly energetic and coherent at periods between 2.5 and 5 days (Reed et al., 1977). (The inertial period at 8°N is 3.5 days.)

These results of Käse and Olbers (1979) provide a hint that non-linear interactions may not saturate the frequency band f to N in low latitudes. It seems possible that there is not a universal spectrum for internal waves in low latitudes, but that the spectrum depends on that of the surface winds.

The most energetic fluctuations, in the vicinity of 8°N 22°W in the upper 200 m, were associated with inertial oscillations, which had velocities of about 15 cm s^{-1} and a downward energy flux. Semi-diurnal internal tides, which had currents of 6 to 8 cm s^{-1}, had no detectable vertical transport of energy (Perkins and Van Leer, 1977). These amplitudes are to be compared with that of the eastward North Equatorial Current which, near the surface, had a speed of nearly 40 cm s^{-1} (Halpern, 1979) and at depths down to 200 m had speeds of 20 to 30 cm s^{-1} (Perkins and Van Leer, 1977).

B. Equatorial Waves

According to Figure 2, freely propagating waves are possible at all frequencies, equatorward of about 5° latitude. Poleward of this latitude there is a frequency band of forced waves, which separates Rossby waves and inertial waves and which is associated with low energy levels. Hence, equatorward of about 5° latitude, we would expect this valley in the energy spectrum to disappear. Presumably the spectral peak at the inertial period will disappear too. Figure 8 shows energy spectra based on the longest existing time series (80 days) made during GATE. Spectra from 10°N, but not those from 5°N and 0°, are seen to have a pronounced inertial peak. At frequencies lower than inertial, the energy tends to increase with decreasing latitude. This is presumably so because a free-wave response is more energetic than a forced-wave response.

The waves observed extra-equatorially propagate latitudinally (except near their turning latitudes). For example, waves with a period of 2 days propagate freely between 15°S and 15°N and in particular propagate freely across the equator. In addition to these waves, there exist close to the equator waves with a modal latitudinal structure (equatorially trapped waves). The question then is whether measurements close to the equator will show a dominance of latitudinal modes or a dominance of latitudinally propagating waves. The answer will depend on the frequency under consideration. It is improbable that fluctuations with short periods (a few hours for example) will have a distinctive structure near the equator. The same is likely to be true of very low-frequency Rossby waves that can propagate freely between high-latitude circles. Figure 9 shows cross-spectra between the velocity components measured at a depth of 300 m at 1 $1/2^{\circ}$N and 1 $1/2^{\circ}$S along 28°W. The motion is coherent at periods longer than 10 days; the phase indicates that the meridional velocity component is symmetrical about the equator, the zonal component anti-symmetrical. This figure suggests that the high-frequency cut-off for equatorially trapped waves is of the order of 10 days. The 80-day time-series is of insufficient length to determine a low-frequency cut-off. Indications are that the kinematics of the low-frequency oscillations correspond to a superposition of linear equatorially trapped waves (Weisberg et al., 1979).

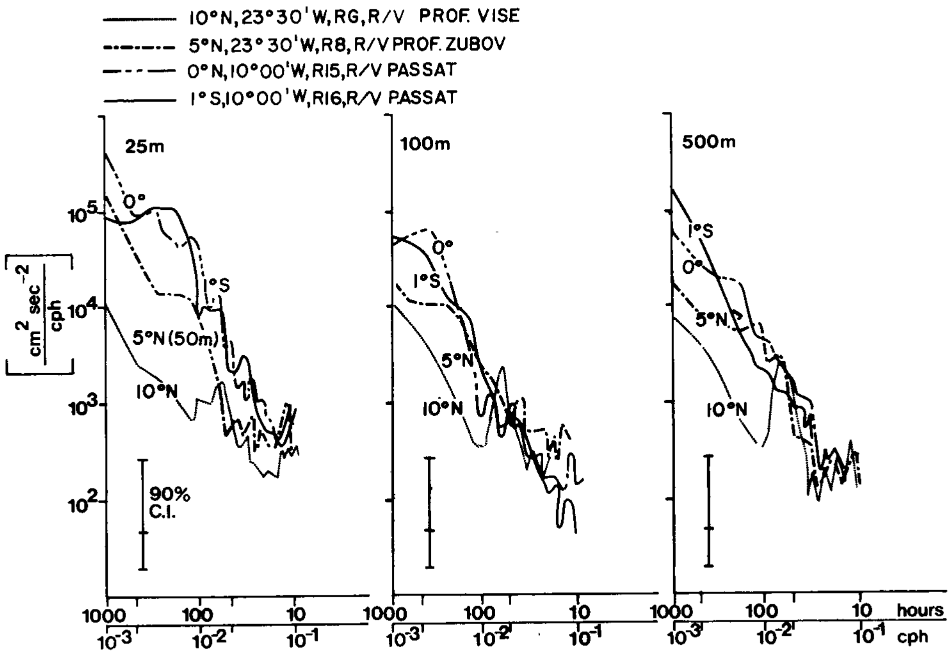

Fig. 8. Kinetic energy spectra of ocean currents at a number of locations (after Bubnov et al., 1979).

A time-series of surface wind measurements during GATE has only recently become available (Krishnamurti and Krishnamurti, 1979). It has therefore not been possible yet to determine whether oceanic and atmospheric fluctuations near the equator are correlated. If they should be correlated then the spectral properties of equatorial waves in the Atlantic and Indian Oceans are likely to be different because the forcing functions are different. The following differences between fluctuations observed in the equatorial Indian Ocean (Luyten, private communication, 1978) and those observed during GATE are noteworthy. At 28°W on the equator, the kinetic energy spectrum for the meridional velocity component (v) has substantially higher energy levels than the corresponding one for the zonal velocity component (u) at periods longer than about 5 days. On the equator in the Indian Ocean at 55°E, on the other hand, u and v have comparable energy levels. Low-frequency fluctuations in the equatorial Atlantic, which have a predominantly downward energy flux and westward phase propagation, have larger coherence scales than fluctuations in the equatorial Indian Ocean. Energy appears to be distributed over a larger number of vertical and latitudinal modes in the Indian Ocean than in the Atlantic Ocean. Whether these differences are attributable to differences in the forcing functions (primarily the surface winds) remains to be seen.

Waves near the equator are propagating through mean currents with considerable latitudinal and vertical shear. The USSR current measurements at 10°W on the equator were sufficiently dense in the vertical to give clues concerning the effect of mean currents on the waves. Weisberg's (1979) analysis of this data shows that fluctuations are particularly energetic, and that the zonal coherence of these fluctuations is very high at depths where the mean zonal flow has maxima (such as the core of the Undercurrent, and its secondary maximum at 200 m). The strong latitudinal shear of the currents complicate the latitudinal phase properites of equatorially trapped waves, according to Weisberg.

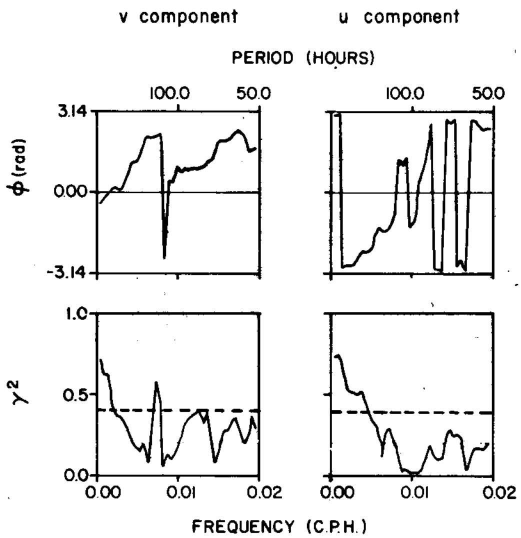

Fig. 9. Phase and coherency squared for currents observed at the 300 m level between 1.5°S and 1.5°N along 28°W (after Weisberg et al., 1979).

It appears that the waves and mean flow interact energetically. The sign of the horizontal velocity component Reynolds stresses indicates that the fluctuations remove kinetic energy from the mean flow (Weisberg et al., 1979). Crawford and Osborn (1979) believe the 16-day meanders of the Undercurrent, to be described below, to be an important sink of mean kinetic energy.

Waves with periods between 3 and 5 days are of particular interest because there is a possibility that they may be resonant. In the equatorial Pacific Ocean there are, at these periods, pronounced peaks in spectra of sea level fluctuations. Wunsch and Gill (1976) have proposed that these peaks correspond to the gravest equatorially trapped inertia-gravity waves that have zero zonal group velocity, and that have the vertical structure of the first baroclinic mode. It follows that random forcing will excite these resonant modes. There happen to be energetic atmospheric waves, with the same temporal and spatial scales, in the equatorial Pacific, but Wunsch and Gill (1976) argue that their presence is not essential for the existence of the oceanic waves. Similar atmospheric waves exist in the equatorial Atlantic but here they appear to be entirely responsible for the energetic equatorially trapped inertia-gravity waves with periods between 3 and 5 days. These waves were non-resonant because they had a downward energy flux. The extremely rough ocean floor, due to the mid-Atlantic ridge, prevents the establishment of a standing first baroclinic mode. Curiously, the amplitude of the velocity fluctuations of these non-resonant waves in the Atlantic is substantially larger that the estimated amplitude for the Pacific. (Wunsch and Gill had sea level data only.) Current measurements are necessary in the Pacific to resolve this matter.

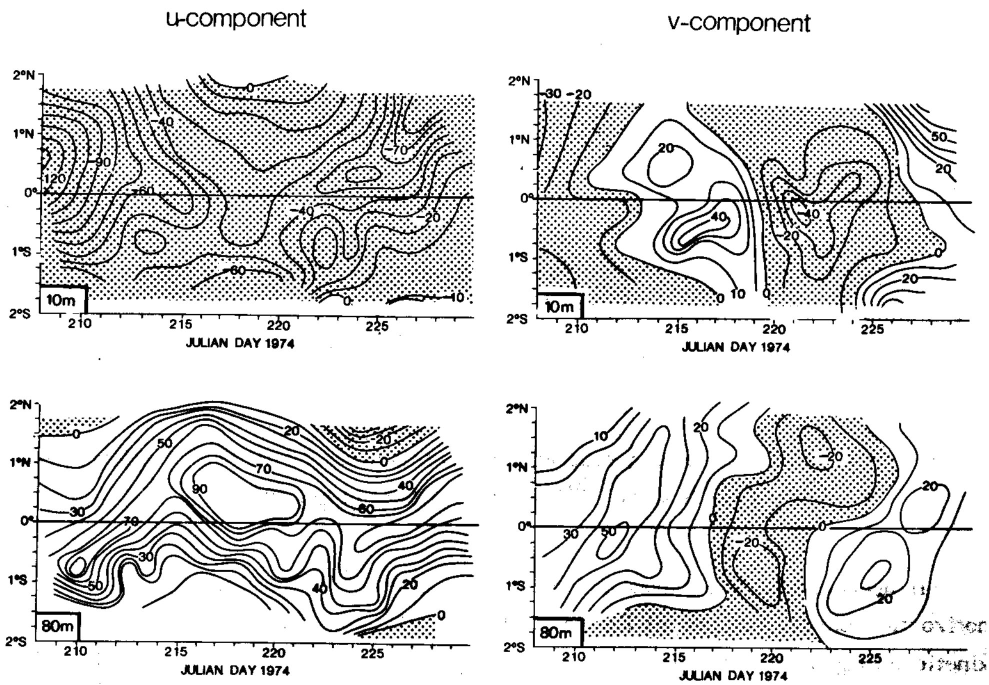

Fig. 10. Latitudinal and temporal fluctuations of zonal and meridional velocity components along 28°W (after Düing and Hallock, 1979).

C. Sixteen-Day Fluctuations in the Upper 200 Meters

Figure 10 shows time-latitude plots of the zonal (u) and meridional (v) velocity components as measured along 28°W and 29°W over a three-week period towards the end of July and in early August, 1974. The most striking feature is the meandering of the Equatorial Undercurrent about the equator. Note that the large values of u north of the equator are preceded by northward flow across the equator and large values of u south of the equator by southward flow. A similar pattern is more difficult to discern in the surface layers because the flow there is not symmetrical about the equator. From these figures the period of the oscillations is estimated to be between two and three weeks.

Spectra of 82-day time series of the meridional velocity fluctuations in the surface layers on the equator at 10°W have a strong peak at a period of about 16 days. Time-series measurements, at fixed points, of fields that are symmetrical about the equator - the zonal velocity component u and temperature T at the depth of the core of the Undercurrent for example - show twice this frequency as the flow meanders back and forth past the observation point. Vertical phase differences are evident in Figure 10: the u fluctuations in the surface layers lead those at the depth of the core of the Undercurrent by about one week. Time-depth plots of the v-components show an upward phase propagation of about 50 m day^{-1} (Düing et al., 1975). The fluctuations appear to be limited to the upper few hundred meters: there is a vertical decrease of amplitude by a factor of about four between the surface and 200 m.

Bubnov et al. (1979) have recently produced daily meridional sections for the three Phases of GATE along 23°30'W, based on current meter moorings, each equipped with ten current meters distributed from the surface to 1000 m. Spatially, this data set is of similar density as the current profiling data set obtained by R/V ISELIN along 28°W. On the basis of these data sets Düing and Hallock (1979) estimate the most likely range of values for the east-west wavelength to be from 1750 km to 1980 km. They also show that coherent westward propagation along the equator over the entire distance from 10°W to 29°W is highly probable. Weisberg (1979) came to the same conclusion based on moored current meters at 10°W, 26°W and 28°11'W. The estimates of the scales of these fluctuations are necessarily approximate, since the time-series are insufficiently long and spatial resolution along the equator is inadequate. To make matters worse, there is the energetic background spectrum of higher-frequency fluctuations described in Section 4(B) above, and an energetic 30-day oscillation to be discussed below in 4(D). The time-series measurements of currents on the equator at 28°W, 26°W and 10°W give more information about the scales, but only at a depth of 200 m where the 16-day oscillation was much weaker than in the surface layers. Weisberg (1979) estimates the zonal wavelength of the 16-day wave at a depth of 200 m to be approximately 1200 km. This is substantially less than the earlier mentioned value of about 1800 km. The difference between the estimates for the wavelength could be regarded as the range of possible errors. There is, however, also a possibility that the meanders of the Undercurrent described here are associated with a phenomenon trapped in the surface layers and that measurements in the deeper ocean are of the background 16-day noise.

Energetic fluctuations such as those described above could be either atmospherically forced or could be due to instabilities of the mean currents. We discuss these possibilities in turn.

Krishnamurti and Krishnamurti's (1979) analysis of surface winds during GATE reveals the existence of a westward propagating atmospheric wave with a period of about two weeks and a wavelength of about 4000 km. These atmospheric fluctuations will excite oceanic waves with the same zonal and temporal scales. The gravest wave to be excited will be a Rossby-gravity wave. Hallock (1977, 1979) has suggested that the 16-day meanders of the Undercurrent correspond to this atmospherically forced Rossby-gravity wave. There is, however, a discrepancy between the wavelengths of the atmospheric and oceanic waves. Perhaps the observed oceanic wave is resonant and not very sensitive to the details of the forcing function: this possibility needs to be explored further. Another possibility is that the observed waves, though only weakly unstable (as shown by Philander, 1978), grow rapidly when the forcing function is particularly energetic at comparable scales. Calculations of Reynolds stresses show that the 16-day fluctuations do indeed draw kinetic energy from the mean flow (Crawford and Osborn, 1979). Again, additional theoretical studies are necessary.

D. One-Month Fluctuations Near the Equator

The most energetic fluctuation observed during GATE appears to have been a westward propagating undulation with a period of about a month and a wavelength of approximately 1000 km. In the surface layers it is most intense in the neighborhood of 3°N, the boundary between the westward South Equatorial Current and the eastward North Equatorial Countercurrent (Fig. 1). Fluctuations on a similar time scale at 0°, 10°W are found to be deep-reaching, having a maximum in v and a minimum in u. Only the three month time-series records (current measurements along 10°W, 23.5°W and 28°W; the 100-day sea surface temperature data set) provide information about this low-frequency wave. Since GATE, however, an array of moorings with current meters below the 500 m level has been maintained in the neighborhood of 0°,4°W. The dominant signal in these long records is the same 30-day (1-month) wave (Weisberg et al., 1979a). Oscillations with practically the same scales have also been observed in the eastern equatorial Pacific (Legeckis, 1977; Harvey and Patzert, 1976; Wyrtki, 1978) so that they appear to be in common in the equatorial oceans.

The fluctuations in the surface layers of the ocean can be explained as an instability due to the large latitudinal shear between the South Equatorial Current and North Equatorial Countercurrent (Philander, 1978). This explains the large amplitude near 3°N.

The unstable waves in the surface layers will radiate equatorially trapped waves, specifically Rossby-gravity waves, into the deep ocean. The 30-day waves measured at depth do indeed satisfy the dispersion relation for Rossby-gravity waves (Weisberg et al., 1979a).

From further analysis of GATE data it may be possible to determine the structure (latitudinal and vertical) of the unstable wave (the forcing function) in the surface layers. If the response of the deep ocean can simply be related to events in the surface layers then the implication would be that non-linear wave interactions do not play as dominant a role in determining the spectrum of equatorial waves as they do in determining the spectrum of extra-tropical waves.

5. MIXING PROCESSES

The surface layers of the ocean are well-mixed and nearly homogeneous. Figure 3 shows that the depth of this mixed layer has considerable latitudinal variation: it is very shallow near 10^{o}N and at the equator, and is deepest in the neighborhood of 4^{o}N. Because of the strong zonal component of the trade winds, which presumably drive warm surface waters to the west, the mixed layer is considerably deeper near the South American coast than near the west African coast. In mid and high latitudes mixing associated with the passage of intense atmospheric storms plays a crucial role in maintaining the mixed layer. In the tropics there are no comparable atmospheric disturbances (save sporadic hurricanes and typhoons). Furthermore the thermocline below the mixed layer is much more strongly stratified in the tropics than in higher latitudes. As a consequence, the passage of the relatively weak atmospheric disturbances - primarily the 3-to 5-day waves - have little effect on the mixed layer. Large changes in the structure of the mixed layer occur on fairly long time-scales; Figure 6 shows changes over the course of GATE. These changes are associated with changes in the entire tropical current system. It follows that models of the mixed surface layer in the tropics cannot be one-dimensional but must take advection by large-scale currents into account (Halpern, 1979; Clarke, 1978). It is clear from Figure 3 that a model that simulates the latitudinal variation in mixed layer depth between 10^{o}N and 4^{o}N must necessarily simulate the North Equatorial Countercurrent. Near the equator the situation is even more complicated because there the ocean is sensitive to fluctuations in the local winds (which induce upwelling). Furthermore, the variability of the surface currents (which at times bring the Equatorial Undercurrent to the surface) has a strong effect on the sea surface temperature.

In addition to the mixed layer at the surface, there is a sub-surface layer of minimum stability, referred to as the thermostad, below the core of the Equatorial Undercurrent. Along 28^{o}W it extends from 7^{o}N to 5^{o}S, is about 100 m thick, and has a mean vertical temperature gradient of $.01^{o}C\ m^{-1}$ (Huber et al., 1978). In an eastward direction, along the equator, this thermostad deepens and becomes warmer and more saline due to the entrainment of fluid from the core of the Undercurrent (Katz et al., 1979). Presumably the vertical shear of the Equatorial Undercurrent makes it possible for the mixing processes to be maintained within 100 km or so of the equator. The thermostad, however, has a larger latitudinal width from which one could infer divergent (poleward) flow at the depth of the thermostad.

In numerical (or analytical) models of tropical phenomena, mixing processes are usually parameterized with constant coefficients of eddy diffusivity. For the mixed surface layer near 8°N 23°W Halpern (1979) estimates a vertical eddy viscosity of about 27 $cm^2\ s^{-1}$. During a wind event it increases to 80 $cm^2 s^{-1}$. Katz et al. (1979) estimate that both above and below the core of the Equatorial Undercurrent the eddy diffusivity is of the order of 1 $cm^2\ s^{-1}$ and the eddy viscosity probably ten times larger. Crawford and Osborn (1979) independently estimate the eddy viscosity to be 10 $cm^2\ s^{-1}$ on the basis of direct turbulence measurements. For a lateral eddy diffusion coefficient at the level of the core of the Undercurrent, Katz et al. (1979) assign a value of $2 \times 10^7\ cm^2\ s^{-1}$.

6. SUMMARY

The principal components of the current system in the tropical Atlantic Ocean are the zonal currents shown in Figure 3. (This figure also shows the depth and latitude dependence of the mean hydrographic fields as measured during GATE.) The most important eastward currents are the North Equatorial Countercurrent between 4°N and 10°N approximately - the associated slope of the density field indicates a geostrophic balance - and the Equatorial Undercurrent with which is associated water of exceptionally high salinity. The principal westward currents are the North Equatorial Current north of 10°N and the South Equatorial Current south of about 3°N. The main longitudinal gradients are in mixed layer (or thermocline) depths, and sea surface temperature. The values of both these parameters decrease in an eastward direction (towards the African continent) along a latitude circle.

Superimposed on the mean conditions is a continuous spectrum of variability. Very high-frequency (turbulent) fluctuations are responsible for the mixed layers which are found at the ocean surface, and below the core of the Equatorial Undercurrent. In the energetics of mixed surface layers in the tropics, advection plays a central role but intermittent mixing by large-scale atmospheric storms does not. Hence one-dimensional models of the mixed layer are inadequate in the tropics. (In higher latitudes, by contrast, the processes that influence the mixed surface layer seem to be mainly local.) The layer of minimum stability (thermostad) below the core of the Undercurrent is due to the large vertical shear of the mean currents there. Since its latitudinal extent exceeds that of the region with high vertical shear, there probably is divergent (poleward) flow at this depth.

In the neighborhood of 8°N 22°W the fluctuations with periods less than the inertial period are inertia-gravity waves. These fluctuations have two unusual features. At periods between 10 min and 1/2 hr the oscillations are exceptionally energetic and correspond to the gravest surface-trapped mode. The prominence of this mode must be due to the shallowness of the thermocline at this location. A second unusual feature is the correlation between the amplitudes of inertial waves below the extremely strong thermocline and atmospheric disturbances. It is surprising that non-linear interactions do not destroy this correlation. Kinetic energy spectra have a pronounced peak at the inertial period. At slightly longer periods energy levels decrease sharply before increasing again as the frequency decreases. This spectral shape changes equatorward of about 5° latitude where the inertial peak is weak (or absent) and the spectrum is uniformly 'red'. In this equatorial region, equatorially trapped waves, which are anti-symmetrical about the equator, dominate the spectrum at periods longer than a week. The statistical properties of these fluctuations are sufficiently different from those in the equatorial Indian Ocean to cast doubt on the existence of a universal spectrum for equatorial waves.

Superimposed on the spectrum of variability described above are exceptionally energetic equatorial fluctuations with periods of 3 to 5 days, 16 days and 30 days. The 3-to 5-day oscillations are equatorially trapped inertia-gravity waves which are excited by atmospheric disturbances with the same period. They propagate downward into the deep ocean. The 16-day waves are confined to the surface layers and cause the Equatorial Undercurrent to meander about the equator with a wavelength of about 1800 km and a westward phase speed. The cause of these waves is unclear, but fluctuations of the tradewinds, with a period of about 2 weeks, may play a role in their excitation. Waves with a period of a month and a wavelength of 1000 km are, in the surface layers, most energetic in the neighborhood of 3°N where the shear between the Countercurrent and South Equatorial Current is large. (The waves are probably due to a shear instability.) The waves in the surface layers radiate equatorially trapped Rossby-gravity waves into the deep ocean.

Further analysis of GATE data that should prove interesting concerns the relation between oceanic and atmospheric variability. Are oceanic waves near the equator correlated with local atmospheric fluctuations? If the answer should be yes, then the implication is that the spectrum of equatorial waves is determined by the local forcing function and is not universal. In that case, the spectra should have considerable geographic variability. Detailed comparisons of spectra from different longitudes (including different ocean basins) will therefore be valuable. Since the non-equatorial wave spectrum is believed to be universal one could also ask at what latitude the change to a non-universal spectrum occurs.

At lower frequencies, the most important result from GATE is the correlation with zero lag between the slope of the equatorial thermocline in the western Atlantic and the intensity of the westward wind-stress, on a seasonal time scale. This result implies that on this time-scale, the waves that effect the adjustment of the ocean to changing winds propagate so fast that there is practically no phase lag between the forcing and the response. The change in the slope of the thermocline has a spatial scale comparable to the width of the basin. Hence the time it takes waves to propagate across the basin must be at most comparable to the seasonal time scale in the Atlantic Ocean. The Pacific Ocean is about three times wider than the Atlantic; hence, in the Pacific Ocean one would expect a correlation between changes in the wind-stress and basin-wide density gradient only on inter-annual time scales. Barnett (1977) does indeed find that the difference in sea-level between the eastern and western equatorial Pacific is correlated with changes in the zonal wind-stress at inter-annual time scales, with practically zero phase lag. Furthermore, Los Niños, which occur interannually in the Pacific Ocean, seem to have a seasonal counterpart in the Atlantic (see Section 3). Studies of low-frequency, large-scale variability in each of the three equatorial ocean basins will be highly complementary.

One would next like to know whether the in-phase correlation between low-frequency oceanic and atmospheric variability is confined to equatorial regions only. Is there perhaps such a correlation for the North Equatorial Countercurrent too? Figure 6 shows that this current changed substantially during GATE. Attempts to correlate these changes with the variability of the winds will be most valuable.

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