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Response of the western Equatorial Atlantic Ocean to an annual wind cycle

by Eli Joel Katz and Silvia Garzoli

ABSTRACT

From the Global Weather Experiment in 1979, four one-year data records are juxtaposed which simultaneously describe the wind stress, thermocline depth, surface current and transport of the Equatorial Undercurrent in the western equatorial Atlantic. The observations are directly compared to predictions derived from a zonally wind-forced nonlinear model of Philander and Pacanowski (J.G.R., 85, 1123–1136, 1980). The correspondence between observations and model suggests that the oceanic cycle of the region can be reasonably described by two sequential events: the response to an everywhere uniform and constant zonal wind which begins in mid-May and the response to the cessation of that wind after January. The response to the wind includes a 40 m deepening of the thermocline in the west, an initially westward, but soon near-zero, surface current and an eastward subsurface transport which rapidly levels off. The response to the relaxation of the wind finds the thermocline rising, the surface current becoming eastward, and the undercurrent transport first increasing to its annual maximum and then decreasing to its minimum just after the wind resumes and the cycle presumably repeats.

1. Introduction

The upper equatorial ocean is remarkable for its relatively quick response to changes in atmospheric forcing. In the monsoonal Indian Ocean, the forcing is itself both strong and rapidly changing, and the resulting circulation is unique to that ocean. The Atlantic and Pacific Oceans share a more modest wind forcing but differ significantly in basin width. The result is a similar mean circulation pattern derived from a similarity in the mean annual wind but differently modified. The Atlantic, less than a third as wide, responds to seasonal changes with little or no memory of interannual changes, while the Pacific is beset by a spatially less homogeneous wind field and burdened by a longer memory.

The distinctions noted above result in the need for different sampling strategies in each of the equatorial oceans. In the Atlantic, the first emphasis should be on year-long measurements, yet almost nothing of that duration has been available.

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Figure 1. Vertical profiles of temperature and zonal velocity at 28W. Velocity is relative to 500 m. Both variables are short time-averages observed simultaneously. Profiles were selected from meridional sections (stations every 25 n.m.) as being closest to the high speed core of the undercurrent.

What is known about the annual response is therefore mostly interpretative. For example, it was deduced that the zonal pressure gradient force acting eastward along the equator is strongest in the boreal summer/fall and weakest in the spring from only the barest few synoptic observations (Katz et al., 1977) and from an historical analysis of the thermocline displacement (Merle, 1980). This cycle had not, however, actually been observed during the course of any one year.

In 1979 the Global Weather Experiment (FGGE) stimulated a number of loosely correlated national oceanographic programs in the equatorial Atlantic. Some of the resulting data span the entire year. An earlier study (Katz et al., 1981) combined some of these data to describe the annual variation of the transport of the Equatorial Undercurrent. The results compared favorably with model predictions of oceanic response and we were encouraged to extend the study to include additional parameters which were observed during the same time period.

The upper equatorial ocean is composed of two distinctive layers. This can be seen from the vertical profiles of temperature and zonal velocity taken during FGGE at 28W and through the core of the undercurrent (Fig. 1). The thermal (and density) profile of the upper waters is characterized by a substantial mixed layer overlaying a strong, thin thermocline. The depth of these layers varies. The zonal
Figure 2. Location of the observations. Four inverted echo sounders (Alice, Branca, Eliana and Flavia) are indicated. The triangle locates St. Peter and St. Paul rocks. The shaded box delineates the area of the meridional current profiling sections.

velocity is not uniform across the two layers. A maximum eastward velocity is found in the thermocline and it locates the core of the undercurrent (a distinctly saline water mass in this region most of the year). In the mixed layer, the surface current varies during the year in both magnitude and direction and is separated from the undercurrent by a shear zone at the base of the mixed layer.

The zonal wind in the equatorial Atlantic, excluding the Gulf of Guinea, decreases at the beginning of the year, increases in the late boreal spring and then is sustained throughout the rest of the year (e.g., Fig. 3, Düing et al., 1980). The classical way of modeling this wind cycle has been to consider first the result of initializing a wind field at time zero and describing the development of the steady state response and, secondly, relaxing the wind after steady state is achieved and then describing the system's decay.

Cane (1979, 1980) has recently investigated the ocean's response to such forcing using a nonlinear model with high horizontal and low vertical resolution. He applied the model to the Indian Ocean to explain the current reversal observed near the island of Gan. Philander and Pacanowski (1980) extended the calculations to a larger basin with higher vertical resolution. Philander (1981) then used the model to investigate how the relaxation of the wind in the central Pacific affects the later stages of El Niño in the east.

In Section 2 we will present new calculations from the Philander and Pacanowski model that can be compared directly to the newly available data. The wind forcing, basin size and the thermal structure of this model resemble the central and western Atlantic.

In Section 3 four data sets are introduced. The first is a year-long wind record from St. Peter and St. Paul rocks (Fig. 2 shows the locations of all the observations). It will be used to parameterize the wind forcing in the western equatorial Atlantic. The second data set consists of four inverted echo sounder records which, when combined, extend over the year. It will be used to parameterize the thermocline
displacement. The last two data sets describe the upper ocean's dynamic response to both local wind forcing and the upstream zonal pressure gradient. They come from profiling current meter sections made by several investigators during the FGGE experiment. The two parameters derived from these observations are the mean near-surface current averaged between 2N and 2S and the transport of the Equatorial Undercurrent.

In Sections 4 and 5 we combine the data and model predictions and then examine the significance of the comparison.

2. Modeling predictions

Equatorial models of the oceanic response to abrupt changes in wind forcing describe the eastward propagation of Kelvin waves from the western boundary across the basin and their return as westward propagating Rossby waves (e.g., O'Brien et al., 1980, Figs. 8-10). The response time of the equatorial ocean depends on the time required for these waves to propagate in from the boundaries and therefore we want to compare the observations with a model which has a thermal structure and basin size adequately resembling the equatorial Atlantic. The significance of the advective terms in the momentum equations has been graphically demonstrated by Cane (1980) and others, and therefore we want to compare with a nonlinear model. The Philander and Pacanowski model (1980) meets these criteria and we will now derive relevant predictions from it. How realistically it includes the diffusion terms can be judged by the results of the comparison.

Their model ocean is a rectangular box, 4800 km wide, 2400 km in meridional extent centered about the equator and 3 km deep. Sixteen grid points are spaced irregularly to define the vertical layering of the ocean. The uppermost grid point is at 12 m. The core of the undercurrent in the thermocline is best represented by the fifth grid point at 112 m. Vertical integration from 0 to 240 m includes half the grid points. The time step is two days.

Motion is forced by imposing a vertical gradient on the zonal velocity at the surface which is proportional to the zonal wind stress. Zero heat flux is imposed by setting the vertical temperature gradient equal to zero at the surface. The model is slowly diffusive (Fig. 3, op. cit.) and this needs to be considered later.

Philander and Pacanowski have described in detail the forced response and relaxation of the upper ocean when the wind is turned on and off. Their results are in a format which highlight the effect of varying the model assumptions (e.g., linear vs. nonlinear response, meridional vs. zonal winds). For our purpose here, we wish to focus on the model results which can be compared most directly to the available FGGE observations. The results are therefore presented in a new format below. They are not new results, only different extractions which are convenient for the comparison.
Spin-up. A zonal wind stress of $-0.5 \text{ dynes/cm}^2$ everywhere is initialized at Day 0 and continued for 300 days. The first 120 days are shown on the LHS of Figure 3. No significant changes occur beyond that period. Six variables are plotted. The zonal velocities at 12 and 112 m at the center of the basin ($L = 2.4$) are shown in Figure 3(a). The near-surface current is initially wind driven and reaches a maximum westward velocity by Day 30. It then decreases to near-zero (Day 75) and thereafter remains small. This equilibrium state of near-zero surface current in the presence of a continuous wind stress is one feature which differentiates non-linear from linear models (e.g., compare Figs. 3 and 8 of Cane, 1980). It results from a momentum exchange with the undercurrent below it which is instrumental in establishing the equilibrium state. The velocity at 112 m is always eastward, counter to the wind stress on the sea surface. It is driven by the zonal pressure gradient force and monotonically increases for the first 60 days. Thereafter it remains relatively unchanged.

The vertically integrated zonal volume transport density (i.e., per unit meridional width at the equator) in the upper 240 m is shown in Figure 3(b). Transport density at the equator has been shown to be a reasonable measure of the meridionally integrated transport of the Equatorial Undercurrent (Firing, 1981; Katz et al., 1981). After a start-up period of westward transport, the net transport becomes increasingly eastward. By Day 75 it has reached its equilibrium range.

The depth of the $15^\circ C$ isotherm at 1.2 and $2.4 \times 10^3 \text{ km from the western boundary}$ is shown in Figure 3(c). The choice of $15^\circ C$ from the model is arbitrary and not critical. It was selected to represent the thermocline depth (rather than a warmer isotherm) because it generally stays subsurface despite strong upwelling in the east. In contrast to the ocean (Fig. 1), the thermocline temperature range of the model was set initially to be 10-25$^\circ C$. To insure that its depth was typical, we compared its pattern with the total heat storage in the upper 240 m and they tracked one another very well (not shown).

The initial depth of the isotherm is 90 m. When the easterly wind begins, the isotherm begins to rise everywhere along the equator due to Ekman divergence. This is arrested by the arrival of a Kelvin wave from the western boundary (Fig. 7(c), Philander and Pacanowski, 1980), first at $L = 1.2$ and later at $L = 2.4$. After 75 days, due to the effect of reflected Rossby waves, the isotherm has reached its equilibrium depth at both locations.

The zonal pressure gradient can be approximated by the upward eastward tilt of the thermocline. The difference in the depths of the isotherm in Figure 3(c) gives this information for the western basin. It is shown in Figure 3(d). The large peak centered around Day 45 can be considered as an artifact of the model which assumes an instantaneous start-up of the wind. In fact, the wind never dies out everywhere and there is no idealized period where the presence of the western
(a) U (112 m)
(b) \( \int_0^{240} U \, dz \)
(c) L = 1.2, L = 2.4
(d) \( \Delta \)
boundary is not being felt in the interior. The progression from initialization to the beginning of the steady state period at Day 60 is likely to be more nearly monotonic.

Spin-down. After Day 300 the wind is abruptly stopped. The RHS of Figure 3 depicts the modeling predictions for each of the above variables during the next 120 days.

When the wind ceases the pressure gradient force is suddenly unopposed by the wind stress, causing an immediate eastward acceleration in the surface layer (Fig. 3(a)). The initially near-zero surface velocity at 12 m strengthens rapidly during the first 20 days. It then more slowly decays as the thermocline slope relaxes, reaching zero by Day 90. The subsurface velocity also increases slightly when the wind stops and then monotonically decreases to zero in 60 days. The net eastward transport density increases during the period of increasing currents and then monotonically, almost linearly, decreases to zero by Day 75.

The depths of the 15°C isotherm quickly converge to an asymptotic value (35 days) at the two locations along the equator; but not to the initial depth of 90 m. This is due to the vertical diffusion in the model. To compensate for this effect, which would occur without imposing any outside forcing, one should discount the slow convergence after Day 90 in Figure 3(c). The RHS of Figure 3(d) would then start with a larger difference and the predicted time for the thermocline tilt to disappear in the western basin would be greater than indicated.

3. Data

Wind stress. Wind was continuously measured from rocks in the western Atlantic (0°50'N, 29°20'W) for one year. This record is being analyzed in detail (S. Garzoli and E. Katz, manuscript in preparation) and it can be stated that the annual cycle of the zonal wind stress reproduces the historical record of the area and is typical of a much larger expanse of the western equatorial Atlantic (as depicted by Hastenrath and Lamb, 1977). Other winds were measured from surface moorings (P. Speth and C. Colin, personal communication) for part of the year and from ships as well. Satellite data from GEOS-East and METEOSAT could also potentially yield a surface wind field for the area, but no one has yet been able to derive a reasonable representation of surface wind from cloud-derived wind vectors over the entire year.
in the Atlantic. The use of the wind data from a single point is therefore neither as bad as it may first seem nor is there any real alternative at this time.

The wind stress components, derived from hourly vector-averaged wind components, and further averaged with a 21-day tapered filter, are shown in Figure 4. The record begins in February 1979 with an already diminishing zonal wind stress which continues to diminish until the end of April. A transitional month is observed in May where the trade wind strengthens and from then until mid-December the zonal stress lies between values of $-0.4 \text{ to } -0.6 \text{ dynes/cm}^2$. The data suggest that 1980 may have seen the trade wind diminish about six weeks earlier than in 1979. This type of interannual variability is normal.

**Thermocline displacement (in the west).** In response to the wind the thermocline will be displaced downward in the west, upward in the east. The result is an eastward acting pressure gradient force which attempts to balance the wind stress. Katz *et al.* (1977) showed evidence that this balance can approximately occur. During the first half of 1979 enough ships were at sea to obtain several estimates of the
Figure 5. Depth of the 20°C isotherm at 40W as derived from IES records. The depth is given by the curves A/B, E and F'. A, E and F are within 14 n.m. of the equator; B is at 1°20'N. F' is the observed F record adjusted from 31W to 40W. The method of adjustment, the procedure used to connect the non-overlapping records and the conversion of the observed travel time delay to depth are discussed in the Appendix.

Gradient, but the observations span a period shorter than we need and they are discontinuous.

Alternatively, we can consider inverted echo sounder records from 31W and 40W (see Fig. 2) which cover the period from mid-January 1979 to mid-February 1980. The conversion of the time series from variation in travel time to thermocline depth is discussed in detail in the Appendix. The resulting composite picture of the thermocline displacement in the western Atlantic is shown in Figure 5 after being low-pass filtered.

Interpretation of event times in the sounder records is somewhat subjective because of energetic displacements at time scales of a month. We find the thermocline at its shallowest during February-April, a deepening is clearly evident from May to mid-July and a marked shallowing occurs first in November, then again in late January of the following year.

Surface current. Between August 1978 and March 1980 twenty-three meridional current profiling sections were made by four independent investigators crossing the equator between 25W and 33W. Four sections are from 1978 and predate the wind record, but they are included as if they were made in 1979. Details of these sections are given in Table 1 of Katz et al. (1981).2 The near surface zonal current (usually about 10 m) of these sections relative to 500 m and averaged between 2N and 2S

2 One additional section is included here. It was made from the OSS Researcher along 27°30'W on 7 March 1980.
Figure 6. Near-surface zonal current relative to 500 m and averaged between 2N and 2S. Data not from 1979 can be identified by comparison with the next figure. Bars indicate the standard error of the mean. Dashed line is interpretative.

(or at least 1N and 1S when the section was short) is shown in Figure 6. This is a most difficult measurement for a profiling current meter and the problem is compounded by the likelihood of energetic short-duration changes in the surface current due to local wind changes. Meridional averaging reduces the effect of these uncertainties.

The resulting average zonal current throughout the year is seen to be close to zero. Strong zonal current was observed during only two periods. From mid-June to mid-July we find an average westward flow of mean speeds between 15 and 55 cm/s. In late February and early March (1980) we find two extremes of non-zero flow, 35 cm/s eastward (observed twice) and 40 cm/s westward. With some license we have drawn a dashed line on Figure 6 to suggest an annual signal which ignores the latter data point. This choice is based on the history of reports of strong eastward currents (one knot) along the equator in the western Atlantic in the spring (Katz et al., 1981, cite the references) from both ship drift reports and a modest number of oceanographic observations by floats and moored current meters. This is not intended to totally discount the late February observation, which should be seen as a warning of possible strongly intermittent flows.

We therefore interpret the record as suggesting strong (but intermittent) eastward flows in February and March, diminishing to zero in May, continuing to the strongest westward flows by mid-July which quickly relax in August and remain near zero until February.
Figure 7. Volume transport of the Equatorial Undercurrent (after Katz et al., 1981). Transport is relative to 500 m and limited to within the 20 cm/sec eastward isotach above 200 m. Open circles are data from 15, 18 and 22W; the rest are from between 25 and 33W. Solid line is interpretative.

Equatorial Undercurrent. The same twenty-three sections discussed above, plus another three farther to the east, have been uniformly analyzed for eastward transport (within the 20 cm/s isotach and again relative to 500 m) by Katz et al., 1981. Figure 7 reproduces their Figure 3 with the additional section mentioned above included. The transport was computed by meridionally integrating a Gaussian least-squares fit to the transport per unit width observed at each station. A range of values from 10 to $44 \times 10^6$ m$^3$/s were reported, with the average equal to $21 \times 10^6$ m$^3$/s.

The solid line offers an interpretation of the transport variation during the year. It basically mirrors the surface flow. Low values of transport occur at the time of strong westward surface flow, and the one instance of very large transport occurs at the time of strong eastward surface flow. Despite the fullness of the curve as drawn, it is clear that we are unable to establish how large or how persistent the transport is during the latter period.
4. Data synthesis and comparison with the model

Figures 4-7 are redrawn in Figure 8(a) on a common time axis. The wind and inverted echo sounder records have been subjectively smoothed to remove scales of
less than a month. Alongside this data synthesis, in Figure 8(b), are shown the predictions derived from the model. The wind stress cycle is fixed first, as discussed below. The results shown in Figure 3 are then transposed in time so that Days 0 and 300 coincide with the indicated wind events.

From the wind record we can identify two time lines: the month of May when the wind stress abruptly changes from near-zero to its full value range of $-0.4$ to $-0.6$ dynes/cm$^2$, and January (in 1979, but in December 1979 for the 1980 season) when the wind begins to relax slowly. We model this wind field, in Figure 8(b), by an abrupt start-up in mid-May and an abrupt cessation at the end of January. The original assumption of $-0.5$ dynes/cm$^2$ in the model is coincidently a good average of the stress during the windy season at 30W during 1979, but the agreement is not overly significant. Winds measured farther to the west (east) would surely have yielded a higher (lower) stress. More important, but less certain, is the zonal variation of the "event" times of wind decay and build-up. These events are determined by the meridional migration of the Inter-Tropical Convergence Zone. The latter is nearly parallel to the equator and therefore the variation in the time of wind onset and relaxation along the equator (in the western ocean) is not expected to be large.

The depth of the thermocline in the western basin is compared for observation and model in the second panels of Figures 8(a) and (b). The data show a deepening of the thermocline from May through July. The difference in February of the two years is not understood, but the slight lead in the deepening of the thermocline before the wind is established may simply result from the wind being recorded 10° east of the thermocline response. The model shows an initial shallowing of the thermocline when the wind begins which we do not necessarily expect to see in the data (where the wind never entirely ceases). After this initial phase, the model predicts that an equilibrium deep-position of the thermocline will be attained in August. The timing, as well as the total displacement (about 40 m), are in good agreement with the observations. The observed thermocline remains relatively deep until next January. There is a slow shallowing drift in the data which was paralleled in the model, but the latter was discounted due to vertical diffusion. The drift in the data may simply reflect a slow reduction in mean wind stress over the western basin. The model's prediction of a return to a shallow thermocline in February is suggested at the tail end of the inverted echo sounder data.

The similarity between the observed and predicted surface velocity (third panels from the top) is particularly strong. We recall that the observations are an average from 2N to 2S along a meridian. The averaging is necessary to smooth the observations. The wind stress imposed on the ocean in the model has no meridional structure. We therefore consider only the predicted zonal current along the equator, where it is a maximum.
Three strong features of the surface current cycle are predicted by the model in the ocean. The model predicts a strong westerly surface current centered around June; it is observed in July. The delay in the observations can be attributed to the non-zero initial condition in the ocean. The important prediction of the subsequent relaxation of the surface current one month later is very clearly seen in the data. Finally, the prediction of the development of a definite eastward surface current in February/March is seen in the data (not unambiguously). Though the present evidence for this eastward current may be less than convincing, it is supported by many earlier observations of the "surfacing" of the undercurrent in the western Atlantic at this time of year.\(^3\)

The total eastward transport is compared with the net transport density in the lowest panels. As above, this difference in quantities plotted is an attempt to make the more meaningful rather than the most direct comparison. Transport mirrors the surface current with one exception: in both observation and model, minimum transport precedes the maximum westward surface current by about one month.

It is clear that the wind resumes before the ocean fully relaxes. Otherwise the undercurrent would vanish, an occurrence never reported for the Atlantic, but nonetheless possible. The model predicts that the transport will reach its equilibrium level by August. It apparently does. The subsequent steady transport is observed. With the cessation of the zonal wind, the model predicts a modest (1/3) increase in net transport, while the data suggest that the eastward transport can double for a short period. Both increases can be traced back to the same cause: the appearance of the eastward surface flow, rather than a large increase in subsurface velocity (see Fig. 3(a) and Fig. 3 of Katz et al., 1981). The 75 days predicted by the model for the net transport to go to zero after cessation of the wind roughly corresponds to the duration that the wind stress is relaxed (though not actually zero). That the Atlantic does not have adequate time to spin entirely down may be attributed to the residual wind stress causing the ocean to relax more slowly than predicted.

5. Conclusions

Three time series describing significant aspects of the low-frequency variation of the upper ocean were compared with a theoretical model. The main features of each variable, including transitional times after a change in wind stress, were at least qualitatively predictable. This agreement leads us to conclude that an on/off steady uniform zonal wind is an adequate forcing function to explain many key responses in the western equatorial Atlantic. Such a simplification is unlikely to be

\(^3\) The observations of Leetmaa and Stommel (1980) in the western Indian Ocean may be relevant. They found a reversal of the surface current about the equator (in May and June, 1975 and 1976) after having been westward in the preceding months. The current reversal followed a relaxation (but not a reversal) of the westward wind component.
valid in the eastern Atlantic where the meridional wind field is dominant at the same time that it is forced from the west (e.g., Adamec and O’Brien, 1978).

A second simplification the data appear to confirm is that the oceanic response to the wind onset and relaxation are for the most part separable. Some recent attempts to force oceans with sinusoidal wind fields (Cane and Sarachik, 1981; Philander and Pacanowski, 1981) or a model of the observed wind over the Pacific Ocean (Busalacchi and O’Brien, 1980) may not be vital to a physical understanding of the western Atlantic.

Equally important is that the model makes no attempt to include realistic off-equatorial currents by including any meridional variation of the zonal wind. Without wind stress curl, for example, there is no North Equatorial Countercurrent, yet the comparison with the data from the equator survives.

Finally, as noted earlier, a specific nonlinear result, the relaxation of the surface current despite steady wind forcing, has been confirmed by the data.

Some differences between the model and the observations were seen to arise because the wind never completely relaxes for a sustained period. The prediction of an initial rise of the thermocline at onset is not surprisingly unobserved. The rate of decrease of the eastward transport in the later stage of the wind relaxation is more gradual than predicted.

The above conclusions are only as good as the data. It should be patently obvious to the reader that we have stretched a meager and inhomogeneous data set to build up the comparison shown in Figure 8. We trust the reader to treat our conclusions with a healthy skepticism at the same time that he looks upon our efforts with kindness. Plans are presently underway to obtain a more homogeneous and robust data set to study the annual cycle of the oceanic response of the equatorial Atlantic.

Acknowledgments. We are indebted to George Philander and Ron Pacanowski for providing the model calculations and John Bruce for his unpublished XBT data. Edifying discussions with Mark Cane and George Philander are gratefully acknowledged. Profiling current meter data from the FGGE experiment were made available by David Cartwright, Hans Lass and Robert Molinari. Richard Payne supervised the wind recorder program; David Bitterman and Esteban Draganovic engineered the inverted echo sounders. Sarah Rennie programmed the data reduction. Instruments were deployed and retrieved from the R/V Conrad, the R/V Oceanus and the Professor Besnard. We appreciate the cooperation of the Brazilian government and the University of São Paulo. Funding was provided by the National Science Foundation under grants OCE 77-25142, OCE 79-22532 and ATM 81-09197.

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APPENDIX

Interpretation of signals from the Inverted Echo Sounders (IES)

The IES’s were floated 1 m off the ocean floor at depths of 1600 to 4200 m. Details of the design of the instrument are given by Bitterman (1976). A burst sample of thirty-two 10 kHz
Table A1. Location and sampling period of inverted echo sounders.

<table>
<thead>
<tr>
<th>Sounder</th>
<th>Location</th>
<th>Water depth (m)</th>
<th>Sampling period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alice</td>
<td>0°13.5’N 40°20.5’W</td>
<td>1700</td>
<td>20 Jan. ’79 - 23 June ’79</td>
</tr>
<tr>
<td>Branca</td>
<td>1°16.0’N 39°08.3’W</td>
<td>1600</td>
<td>22 Jan. ’79 - 25 June ’79</td>
</tr>
<tr>
<td>Eliana</td>
<td>0°04.2’N 39°44.8’W</td>
<td>4200</td>
<td>26 June ’79 - 3 Oct. ’79</td>
</tr>
<tr>
<td>Flavia</td>
<td>0°02.7’S 30°57.1’W</td>
<td>4100</td>
<td>18 Oct. ’79 - 27 Feb. ’80</td>
</tr>
</tbody>
</table>

Note: Shallower depths are from isolated mid-ocean ridges surrounded by water of greater than 4000 m depth.

Pulses are transmitted once an hour and the two-way travel time after reflection from the sea surface is recorded internally. The standard deviation of a sample is usually between 0.3 and 0.6 ms. The median value is selected for analysis, and with the appropriate assumptions it is reasonable to assume that it has been measured to at least ± 0.1 ms.

The unfiltered time series are dominated by the change in sea surface elevation due to the barotropic tides. Cartwright (1982) has analyzed the records and reports tidal amplitudes of 0.8-1.1 ms (M2), 0.3-0.4 ms (S2, K2), etc. The records also include high frequency oscillations due to large vertical scale inertial gravity waves. Garzoli and Katz (1981) report mean amplitudes of 0.2 ms at discrete frequency bands between 2 and 4 cycles/day.

The four echo sounder records were shown in Figure 5 after filtering to remove high frequency variability (low-pass filter with a half-amplitude of 8 days). Sampling periods are given in Table A1. The time delay is a relative scale and the best that can be done is to assume that one record begins where the previous one ends. Alice and Branca were nearly simultaneous and nearby one another. Their mean delay times were set equal. To convert the resulting travel time series to a thermocline displacement we compute a calibration ratio from historical data, interpret the ratio in terms of a free vertical mode, and then compare the result with direct observations.

**Calibration.** From the historical data set we can derive a relationship between travel time delay and thermocline displacement. For each station we compute a hypothetical two-way travel time from 0 to 500 m, assuming no change in surface elevation (a 10% error which we ignore) or change below 500 m (an assumption which we are unable to substantiate a priori). This is then compared to the depth of the 20°C isotherm, which is characteristic of the thermocline depth in this region.

From the NODC files we found 52 usable stations in a 2-degree square centered about 0N 40W. The data are shown in Figure A1, and a linear regression of computed travel time (t) on the depth (Z) of the 20°C isotherm (chosen to represent the thermocline) yields the ratio

$$\frac{\Delta Z}{\Delta t} = -20.5 (\pm 1.6) \text{ m/ms}.$$

The same exercise for 55 stations centered about 0N 30W yielded $-19.7 (\pm 2.3) \text{ m/ms}$. The variance about the regression is relatively large (S.D. = ± 11 m), but much of this could be due to interpolation between discrete bottle depths, calibration variables and high frequency oscillations.

**Interpretation.** Under certain conditions (e.g., Moore and Philander, 1977) the vertical displacement of a stratified water column can be described by a superposition of free standing
vertical modes. The delay in travel time due to each mode is described by Garzoli and Katz (1981) as a function of the mode amplitude (their Eq. 13). The delay depends on the local density profile and they compute the vertical structure of the five lowest baroclinic modes for average conditions at 0N 40W and 4500 m water depth (their Fig. 4). If we assume one-by-one that the vertical displacement is given by a single mode, then the amplitude of that mode can be expressed as a function of thermocline displacement (for definiteness assigned at the depth of maximum vertical density gradient). The amplitude of the two-way travel time delay from any depth, for a 1 m displacement of the thermocline, is shown in Figure A2 for vertical mode numbers, \( r = 1 \) through 5.

To compare with the earlier calculation, we compute \( \Delta Z/\Delta t \) with \( \Delta Z = 1 \) m and \( \Delta t \) equal to the travel time delay at 500 m in Figure A2. The ratio for each mode is given in Table A2. We note that \( r = 2 \) agrees with the ratio computed from the historical data and conclude that the vertical displacement of the upper ocean as seen by an IES can be modeled by the second baroclinic mode (though not uniquely, of course).

**Table A2. Ratio of thermocline displacement to delay in two-way travel time from 500 m.**

<table>
<thead>
<tr>
<th>Mode No., ( r )</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \Delta Z/\Delta t ), m/ms</td>
<td>-13</td>
<td>-20</td>
<td>-500</td>
<td>+75</td>
<td>+28</td>
</tr>
</tbody>
</table>
Figure A2. Delay in two-way travel time due to a 1 m displacement of the thermocline as a function of IES depth. The calculation is repeated assuming that only the $r^{th}$ baroclinic mode is established.

If we assume that the second mode is established throughout the water column, then Figure A2 gives a recipe for selectively converting travel time delay to thermocline displacement for a given IES depth. From Table A1 we note that the sounders were deployed at depths of about 1500 and 4000 m. At 1500 m there is no substantial difference relative to 500 m; but at 4000 m a 30% difference is indicated in the figure. However, there is no experimental evidence that affirms this assumption and theory (Philander and Pacanowski, 1981) indicates that no single mode is dominant below a depth of a few hundred meters. We, therefore, have not corrected for deeper sounders.

Validation. In Figure 5 travel time delays were converted to changes in displacement using a ratio of $-20$ m/ms. The first three sounders are near 40W, and the assumptions made to combine the records are reasonably accurate. Flavia, however, is at 31W and the annual variation of the thermocline depth is reduced to the east. Computations from the historical data at 30W and 40W suggest that the mean annual displacements at these sites are 30 m and 40 m, respectively. The effect of increasing the relative time delay by the factor $4/3$ ($F'$ compared to $F$) is shown in Figure 5. In discussing thermocline displacement at 40W we have used $F'$.

On three occasions, when retrieving echo sounders, CTD stations were made along the equator between 30W and 40W. Rather than compare directly with the depth of the 20°C
Figure A3. Three comparisons of the composite IES record with hydrographic data. Depth is of the 20°C isotherm at 40W. The IES record is from Figure 5 (A/B, E, F'). Upper panel: With in situ data at the time of retrieval. Middle panel: With XBT sections observed by John Bruce one year later. Lower panel: With hydrographic data from NODC.
isotherm from the station at 40W, we have fitted a regression line to the isotherm depth along the zonal section. In Figure A3 (upper panel) we plotted its intersect at 40W and the expected error of the estimate. This is necessary to reduce temporal variability, which appears as spatial variability in the sections. Even with this approach, the temporal variability problem is apparent when we note that a repeated section in February 1980 gave estimates that differ by 11 m.

The \textit{in situ} data are too few to improve upon our earlier estimate of $\Delta Z/\Delta t$. Instead, we use it only to estimate the absolute depth of the 20°C isotherm by adjusting the fit and then make two additional comparisons to establish a degree of confidence in the interpretation of the IES record. The following year, six XBT sections which cross the equator at about 38W were obtained by John Bruce of the Woods Hole Oceanographic Institution. The mean and standard deviation of the depth of the 20°C isotherm from seven bathythermographs per section between IN and IS are compared to the sounder record in the middle panel. In the lowest panel, the sounder record is compared with the historical data used for the calibration.

We conclude from the three comparisons that the composite IES record is a reasonable representation of the annual cycle of the 20°C isotherm in the region. The amplitude of the annual signal described by the IES is perhaps somewhat smaller than the data shown in the lower two panels, but we cannot definitely establish whether or not this is a calibration problem. It also suggests that the deepening of the isotherm was a month or two earlier in 1979 than in other years. The latter is important in the present study of the ocean response time and it is independent of the calibration question.

REFERENCES


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