Deep western boundary current variability off northeastern Brazil

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Abstract—Observations from a year-long deep current meter mooring at 8°N, 52°W are used to describe the structure and variability of the Deep Western Boundary Current (DWBC) in the tropical Atlantic. The DWBC has a deep core near 4300 m depth, is extremely swift and narrow (35 cm s\(^{-1}\) core speed, 60 km width), and shows a total transport of 22 Sv between 2500 m and the bottom. Approximately 3 Sv of the DWBC transport are carried in near-bottom layers below \(\theta = 1.8^\circ\mathrm{C}\), which appears to be recirculating in the western tropical Atlantic north of about 4°N.

Variations in the DWBC are dominated by large-amplitude vertical displacements of the deep temperature field at 60–70 day periodicities that cause significant changes in the deep stratification and vertical transport structure of the DWBC. These fluctuations appear to be related to periodic surges of cold bottom waters up onto the base of the continental rise that become entrained in the DWBC. While the forcing mechanism for these deep fluctuations remains unexplained, the picture emerging from these data is that of a highly active abyssal layer in the tropical Atlantic.

1. INTRODUCTION

In an earlier paper, JOHNS et al. (1990) described the seasonal structure and mesoscale variability of upper layer currents off northeastern Brazil during 1987–1988 using moored current meter data. One of the moorings used in that study, located at the base of the continental rise near 8°N, recorded strong deep southeastward flows associated with the Deep Western Boundary Current (DWBC). The magnitude of these flows was surprisingly large, with maximum instantaneous speeds of up to 45 cm s\(^{-1}\) in the lower 1000 m and an annual mean speed of nearly 30 cm s\(^{-1}\) at 4300 m depth.

The purpose of this paper is to describe the structure and temporal variability of the DWBC in more detail than in the earlier paper, and to also bring to light evidence of large amplitude vertical displacements of the deep temperature field associated with periodic intrusions of cold bottom waters up onto the continental rise. Certain of these events appear to be related to meandering of the DWBC, while others suggest a different mode of variability involving a vertical compression of the DWBC core layer velocity and density structure. The observations also show a net equatorward transport of approximately 3 Sv (1 Sv = 10\(^{6}\) m\(^3\) s\(^{-1}\)) in the DWBC of waters colder than \(\theta = 1.8^\circ\mathrm{C}\), which contain a mixture of Antarctic Bottom waters that appear to be recirculating cyclonically in the tropical Atlantic north of 4°N.

This paper is organized as follows. In Section 2 we first describe the large scale...
hydrography of the region, focusing on abyssal layers, using data collected during 1987–1989 as part of the STACS (Subtropical Atlantic Climate Studies) program. Other aspects of this data set are described by Molinari et al. (1992, hereafter referred to as MFJ-92). Section 3 contains a description of the current meter time series. In Section 4 we develop a simple model of the mean DWBC structure based on the current meter and hydrographic data, and in Section 5 we describe the variability about this mean structure. Section 6 discusses possible forcing mechanisms for the observed variability.

2. HYDROGRAPHIC BACKGROUND

A large scale picture of the deep circulation and water mass properties of the Atlantic Ocean may be found in the atlases of Wüst (1935) and Wright and Worthington (1970). Antarctic Bottom Water (AABW) flows northward into the Atlantic from its formation region around Antarctica and forms a sloping water mass boundary with the overlying North Atlantic Deep Water (NADW). Northward flow of the coldest of these bottom waters appears to be hydraulically controlled by two major topographic sills in the western basin of the Atlantic, one near 30°S [the Rio Grande Rise, separating the Argentine and Brazil Basins (Hogg et al., 1982)], and another near 4°N [the Ceara Rise, separating the Brazil and Demerara Basins (Whitehead and Worthington, 1982)]. Above this layer NADW flows southward along the western boundary across the equator and into the South Atlantic.

North of the Ceara Rise the vertical water mass gradient between AABW and NADW becomes particularly sharp, as evidenced by a pronounced deep stratification maximum near \( \theta = 1.4^\circ \)C (Fig. 1). Figure 2 shows the topography of deep potential temperature surfaces in the Demerara Basin between \( \theta = 1.4^\circ \) and 2.0°C, based on data from four hydrographic cruises during 1987–1989 (MFJ-92), plus additional station data collected in January 1990 (Molinari, personal communication), and selected GEOSecs stations (Bainbridge, 1981). The \( \theta = 2.0^\circ, 1.8^\circ, \) and \( 1.6^\circ \)C surfaces all show maximum depths within the center of the basin, with isotherms deepening to the north and shoaling toward the western boundary and toward the Mid-Atlantic Ridge. At \( \theta = 2.0^\circ \)C the isotherms slope most steeply on the western side of the basin, indicating southward shear associated with the DWBC, while at \( \theta = 1.6^\circ \)C the isotherms slope most steeply up toward the Mid-Atlantic Ridge, indicating deep northward flow of bottom waters along the western flank of the ridge. The northward deepening of these isothermal surfaces also becomes progressively larger with depth, such that at \( \theta = 1.4^\circ \)C the isotherms are characterized by essentially a monotonic increase in depth toward the north with little cross-basin structure. (Note that the data distribution is quite sparse in the eastern portion of the basin, so that the contours drawn there are rather subjective.)

During the hydrographic cruises several sections were repeated, with the largest accumulation of data along a section extending northeastward from the western boundary between 7 and 10°N (Fig. 3). As can be seen in Fig. 4, the offshore deepening of isotherms associated with the DWBC is concentrated within about 150–200 km of the western boundary, with largest mean slopes in the \( \theta = 2.0–2.5^\circ \)C range (approximately 3000–4000 m). The core layer of lower NADW appears as a stratification minimum near about 1.9°C, underlain by a thin bottom water layer with potential temperatures as cold as 1.3°C. A notable difference between these two sections is the strong reversal in isotherm slopes below about \( \theta = 1.8^\circ \)C that occurs near the western boundary in September 1987. This
implies a shear reversal (and hence southward velocity maximum) within the DWBC somewhere near 4300 m, whereas during September 1988 this reversal occurs both deeper (near $\theta = 1.6^\circ$C) and farther offshore, such that near the western boundary the southward velocity shear appears to extend nearly to the bottom. Absolute velocity profiles also were obtained during these sections at two locations (only one of which was in deep water within the DWBC, see Fig. 3) (Fig. 5. Leaman, personal communication). These profiles show more clearly the differences in the shear structure of the DWBC during these occupations, i.e. the pronounced velocity maximum near 4300 m during September 1987, vs the broad, weakly-sheared core during September 1988, with strong speeds extending from approx. 3700 m to near the bottom. Further evidence for variability in the deep shear structure of the DWBC is found in the moored current meter data, which suggest more dramatic temporal changes than those shown in Figs 4 and 5, as described below.

3. MOORED CURRENT METER OBSERVATIONS

The section shown in Figs 3 and 4 was sampled by a deep current meter mooring deployed from September 1987 to September 1988, at the location indicated in the figures. Current meter observations were collected at nine levels on this mooring. Five of the current meters were located below 1000 m, at nominal depths of 1600 m, 2500 m, 3800 m, 4300 m and 4600 m (see Table 1). All instruments recorded velocity and temperature, while only the three uppermost instruments (at 100 m, 200 m and 500 m nominal depths) recorded pressure. Depth variations of the deeper current meters were approximated...
using a mooring performance program, for later use in correcting the temperature measurements for mooring motion.

Time series of vector velocity below 1000 m on this mooring (Fig. 6) show the DWBC in the three deepest records as a strong and fairly persistent flow to the southeast, except during the first two months of the record where the flow is weaker and even reverses briefly. The 1600 m and 2500 m records show greater variability, and some of the events in these records appear to be coherently related to energetic upper layer current variations seen in this region [see Johns et al. (1990) for a more complete description of the upper ocean variability here]. The DWBC increases from about 3 cm s\(^{-1}\) at 1600 m depth to a maximum of 28 cm s\(^{-1}\) at 4300 m depth, before decreasing again toward the bottom (Table 1). Except for the 1600 m level, the deep mean flows are all within about 5 degrees of the direction of the local isobaths, which run about 120° true.

*In situ* temperatures were recorded by the current meters, with an absolute accuracy of
about 0.01°C. These temperatures were converted to potential temperatures using a regional T–S relation derived from CTD data (Fig. 1). Surprisingly large potential temperature fluctuations occurred deep in the water column (Fig. 7), with peak-to-peak amplitudes of as much as 0.5°C at 4600 m depth (100 m off the bottom). These fluctuations appear as bottom-intensified cold events with characteristic time scales of 1–2 months. In general these fluctuations have a wavelike appearance although the temperature extrema tend to be sharply peaked rather than sinusoidal, particularly near the bottom. Figure 7 also shows equivalent vertical displacement time series for these records, \( dz(t) \), calculated from:

\[
dz(t) = \left[ \frac{\partial \theta}{\partial z} (\theta) \right]^{-1} \cdot d\theta(t) \cdot dt
\]

where \( \frac{\partial \theta}{\partial z}(\theta) \) is the mean vertical potential temperature gradient calculated from CTD data (its structure is analogous to that shown in Fig. 1). As can be seen, the inferred vertical displacements at the two deepest levels are quantitatively similar and show typical amplitudes of 200–300 m. The vertical displacements at the next level (900 m off the bottom) are greatly reduced in amplitude, and correlate well with the deeper levels only during certain of the events. The spectra of these deep temperature time series (and the corresponding vertical displacement time series) are peaked at periodicities of about 60–70 days (Fig. 8).

Inspection of Figs 6 and 7 shows that these cold events occurred only during the last two-thirds of the record when the strong southeastward flow of the DWBC was present at the
Fig. 4. Cross-sections of potential temperature off northeastern Brazil for (a) September 1987 and (b) September 1988. Also indicated are the location of the mooring and the depths of individual current meters.

mooring site. However, apart from this general correlation there appears to be no systematic relationship between temperature and velocity fluctuations during these events. The velocity spectra do not show a similar 60–70 day peak, and the coherence between deep velocity and temperature fluctuations is very small except in a limited band near 10–20 days where colder temperatures appear to be weakly correlated with stronger equatorward velocities (Fig. 7). The relative steadiness of the DWBC during these large temperature events is in fact quite remarkable, and below we attempt to provide some insight into the nature of these events and their relationship to variations in the DWBC structure.
Fig. 5. *Pegasus* profiles of alongshore velocity component below 1500 m at the mooring location, (a) September 1987 and (b) September 1988.

4. A MODEL OF THE MEAN DWBC STRUCTURE

To interpret these events it is first useful to construct a model of the mean DWBC structure at this location. One approach would be to average the available hydrographic sections, although in view of the substantial differences between individual section occupations (e.g. Fig. 5) it is not clear how such an average should best be computed. A further drawback to this approach is that due to the relatively large station spacings these sections may not provide a very well resolved picture of the DWBC structure.

A different approach is to use the direct velocity observations from the mooring, averaged in such a way as to be representative of the mean vertical structure near the axis of the DWBC, and to combine this with some assumption about the lateral structure of the DWBC. This model can then be constrained by forcing it to be consistent with the mean

<table>
<thead>
<tr>
<th>Lat.</th>
<th>Long.</th>
<th>Water depth</th>
<th>Instrument</th>
<th>Depth (m)</th>
<th>Length (days)</th>
<th>East (u)</th>
<th>North (v)</th>
<th>Vector mean</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Mean</td>
<td>S.D.</td>
<td>Mean</td>
</tr>
<tr>
<td>08°30.0'N</td>
<td>05</td>
<td>1600</td>
<td>315</td>
<td>1.3</td>
<td>(6.8)</td>
<td>-2.6</td>
<td>(6.4)</td>
<td>2.9</td>
</tr>
<tr>
<td>52°09.0°W</td>
<td>06</td>
<td>2500</td>
<td>363</td>
<td>5.6</td>
<td>(6.2)</td>
<td>-2.7</td>
<td>(4.9)</td>
<td>6.2</td>
</tr>
<tr>
<td>4740 m</td>
<td>07</td>
<td>3840</td>
<td>363</td>
<td>16.7</td>
<td>(8.8)</td>
<td>-11.6</td>
<td>(7.8)</td>
<td>20.3</td>
</tr>
<tr>
<td></td>
<td>08</td>
<td>4340</td>
<td>363</td>
<td>24.7</td>
<td>(12.1)</td>
<td>-13.0</td>
<td>(7.1)</td>
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<td></td>
<td>09</td>
<td>4640</td>
<td>363</td>
<td>14.7</td>
<td>(8.0)</td>
<td>-8.3</td>
<td>(6.3)</td>
<td>16.9</td>
</tr>
</tbody>
</table>

Velocities are in cm s\(^{-1}\). Direction is degrees from true North.
DWBC transport estimated from hydrographic sections. DWBC transport estimates derived from the available hydrographic sections are given in MFJ-92, and summarized in Table 2.

Using this approach, we take as a model for the DWBC a Gaussian jet of the form:

\[ \nu(x,z) = \nu(z) \cdot e^{-(x-x_0)^2/(2\sigma^2)} \]  

(1)
where $v(z)$ is the observed velocity profile at the axis of the jet ($x = x_0$), and where $x_w$ is the Gaussian half-width of the jet. Once this velocity field is specified, horizontal density gradients can be determined, assuming geostrophic balance, by

$$\frac{\partial \rho}{\partial x} = -\left(\frac{\rho f}{g}\right) \frac{\partial v}{\partial z},$$

(2)

or, in terms of horizontal potential temperature gradients:

$$\frac{\partial \theta}{\partial x} = -\left(\frac{\rho f g}{\partial \rho / \partial \theta}\right)^{-1} \frac{\partial v}{\partial z},$$

(3)

where $\partial \rho / \partial \theta$ can be determined at any given pressure level using the deep $\theta$--S relation (see Fig. 1). A consistent representation of the deep potential temperature field can be constructed by integrating (3) outward from the jet axis, starting from the observed potential temperature profile $\theta(z)$ at $x_0$. 

Fig. 7. Time series of (a) potential temperature (bottom panel) and (b) vertical displacement (top panel) for the three deepest instruments on the mooring. See text for details on the vertical displacement calculation. The vertical displacement time series are offset by 400 m.
Fig. 8. Variance-conserving spectra of (a) alongshore current and (b) potential temperature, ensemble averaged from the three deepest observation levels. Coherence $^2$ (c) and phase (d) are also shown. 95% confidence limits on the auto-spectra are indicated.

<table>
<thead>
<tr>
<th>Source</th>
<th>&lt;1.8°C</th>
<th>1.8–2.0°C</th>
<th>2.0–2.4°C</th>
<th>2.4–2.8°C</th>
<th>2.8°C-bottom</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model</td>
<td>-3.6</td>
<td>-7.6</td>
<td>-6.9</td>
<td>-4.0</td>
<td>-22.1</td>
<td></td>
</tr>
<tr>
<td>Molinari et al., 1992*</td>
<td>$-3.1 \pm 0.5$</td>
<td>$-7.4 \pm 1.2$</td>
<td>$-6.8 \pm 1.1$</td>
<td>$-4.5 \pm 1.0$</td>
<td>$-21.8 \pm 2.1$</td>
<td></td>
</tr>
</tbody>
</table>

* Values correspond to the “French Guiana” section (their Table 2), which crosses the mooring site.
To obtain average $v(z)$ and $\theta(z)$ profiles, we first created daily-averaged profiles of $v$ and $\theta$ by fitting a shape-preserving cubic spline function (Akima, 1970) through the four deepest observation levels, between 2500 m and the bottom. These profiles were then averaged over the last 260 days of the record when strong DWBC flows were present at the mooring site. These average profiles were then used with different values of $x_w$ to create modeled cross-sections of $v$ and $\theta$, from which the total volume transport below 2500 m, as well as transports in different potential temperature classes, were calculated. The total transport was then compared to the corresponding mean value estimated from the available hydrographic sections, to select the most reasonable choice of $x_w$. (The reference level for the hydrographic calculation was, in turn, chosen to be consistent with the current meter observations; see MFJ-92.)

Figure 9 shows the model DWBC cross-section for $x_w = 30$ km, which provided the best agreement with the total DWBC transport below 2500 m from the hydrographic data. The model-derived transports include all contributions inside the 2 cm s$^{-1}$ isotach (chosen as a practical limit since in theory a Gaussian jet has infinite transport). As shown in Table 2, both the total DWBC transport of 22 Sv (below 2500 m) and the breakdown of this transport into selected temperature classes are in good agreement with the hydrographic estimates. We therefore take this model to be a reasonable representation of the DWBC structure at this location. It should be noted that the DWBC in the tropical Atlantic also contains a secondary, shallower core near 1500 m depth that flows southward over the continental slope in approx. 2000 m water depth (MFJ-92), which was not well sampled by

![Cross-section of the model Deep Western Boundary Current with a Gaussian half-width of 30 km. Isotherms are solid lines, isotachs are dashed.](image-url)

Fig. 9. Cross-section of the model Deep Western Boundary Current with a Gaussian half-width of 30 km. Isotherms are solid lines, isotachs are dashed.
this mooring. Therefore we have restricted our analysis and comparison with the hydrographic estimates to the lower 2500 m, spanning the deep velocity core.

The model jet shows core velocities of close to 35 cm s$^{-1}$ centered around the $\theta = 1.8^\circ$C isotherm, with a cold wedge of bottom waters extending shoreward beneath the core to form a sharp temperature front near the bottom. The deep isotherms also widen into a lens of very weakly stratified water between about $\theta = 1.7$ and $1.9^\circ$C shoreward of the velocity core. These features are generally consistent with the properties shown in the hydrographic sections (Fig. 4), except that the spreading of deep isotherms near the western boundary is much more pronounced in the model. The frequent occurrence of almost neutral stratification at the deep current meter levels suggests that this structure is probably real (in the mean), but that it may have been either poorly resolved or perhaps absent during the available section occupations.

Perhaps the most interesting aspect of this model is the relatively narrow width it implies for the DWBC. According to the model, core velocities greater than 10 cm s$^{-1}$ are confined to within about 60 km of the western boundary, which is a substantially narrower representation of the DWBC than provided by the hydrographic data (MFJ-92). It should be emphasized, however, that this is the maximum width of the DWBC that is consistent with both the observed deep velocities and the transport estimates from the available hydrographic sections.

5. DWBC VARIABILITY

To provide a clearer perspective of the DWBC variability we first show an alternate representation of the moored velocity and temperature records (Fig. 10), which emphasizes changes in the vertical shear of the deep currents. One feature that emerges clearly in this picture is the strong correlation between deep temperature fluctuations and the shear structure of the DWBC. During the cold events, the DWBC tends to have a more pronounced core-like structure, with sharper shears above and below the velocity maximum, than during the warmer periods. As noted before, however, these cold events (and the associated core intensification) do not always coincide with changes in the vertically averaged flow speed, or local transport, of the DWBC. In fact, the majority of the cold events during the last two-thirds of the record appears to be largely unrelated to local transport changes.

One anticipated mode of variability is lateral meandering of the DWBC, for which Fig. 9 serves as a useful guide. The simplest form of meandering would be a vertically-uniform lateral displacement of the density surfaces, in which the density field retains its basic shape and no vertical stretching is involved (except near topographic boundaries). An offshore displacement of the DWBC therefore would be characterized by decreasing flow speeds at the mooring site coupled with decreasing stratification and generally cooler (warmer) temperatures above (below) the velocity core. Conversely, an onshore displacement of the DWBC to a position shoreward of the mooring site would be characterized by decreasing velocities coupled with increasing stratification and warmer (cooler) temperatures above (below) the velocity core. Offshore meandering would be expected to account for the largest variability in flow speeds, since on average the DWBC core appears to be very close to the continental rise, so that its capacity for onshore meandering is limited by the topography.

Considering again Fig. 10, it appears likely that during the first 90 days of the record the
Fig. 10. Contoured representation of velocity and potential temperature fluctuations below 2500 m at the mooring site. Panel (a) is the vertically averaged speed of the DWBC between 2500 m and the bottom, lowpassed with 30 day running filter. Panel (b) shows velocity anomalies relative to the instantaneous vertical average in (a), also lowpass filtered. [Thus at each time, the anomaly profile has zero vertical mean, while the full velocity profile can be reconstructed by simply adding the value in panel (a) to the respective anomaly profile in (b).] Panel (c) shows similarly filtered contours of potential temperature. The meandering event (★) and anomalous warm (W) and cold (C) events described in the text are indicated.

DWBC was flowing generally offshore of the mooring site except for the period around day 50–60 when the DWBC meandered onshore, bringing the velocity axis near to, but probably still slightly offshore of, the mooring site. Around day 100 the DWBC again meandered onshore, and from this point onward remained fairly well entrenched over the mooring site as indicated by the large vertically averaged speeds. Of the five cold events...
that occurred after that time, one of them (centered around day 240) was accompanied by a significant reduction in core strength and local transport and could be consistently interpreted as an onshore meander of the DWBC core to a position slightly shoreward of the mooring site.

The remaining four cold events, and the warm events interspersed between them (labeled C and W in Fig. 10), do not bear these same transport signatures and cannot be consistently interpreted as uniform lateral meandering of a quasi-fixed DWBC structure. Obviously, with only one mooring, it is not possible to provide an unambiguous interpretation of these events. However, some insight into the structural changes that accompany these events can be obtained by applying the methodology of Section 4 to the anomalous temperature and velocity profiles observed during these events. To illustrate these changes we composited the four warm and four cold events separately, using 10-day averages centered around the times shown in Fig. 10, and used the resulting mean temperature and velocity profiles to create new cross-sections of the DWBC structure corresponding to the average "warm" and "cold" cases. It is assumed that these profiles are representative of conditions near the axis of the DWBC (due to the consistently large vertically averaged speeds), and that the DWBC width remains nearly constant during these events. It is possible that the DWBC width may vary substantially during these events; however there is no way to determine this from the available data. Therefore, in comparing these two cases we are concerned mainly with changes in the vertical structure and the relative transport contributions carried by different temperature classes, which remain qualitatively unaffected by changes in width.

Results for the composited "warm" and "cold" events, for $x_w = 30$ km (Fig. 11) show similar core velocities of about 35 cm s$^{-1}$, the major difference being in the vertical scale of the DWBC core. Relative to the mean DWBC structure (Fig. 9), the cold event case shows a more compressed velocity core, while the warm event case shows a vertically stretched core. The major difference in the temperature field is the larger shoreward intrusion of bottom waters beneath the DWBC core, with waters as cold as $\theta = 1.6^\circ$C extending up onto the base of the continental rise. Also important is the change in the temperature at the velocity core, from 1.7 to 1.8$^\circ$C in the cold events to 1.8 to 1.9$^\circ$C in the warm events. Associated with this increase in the core layer temperature during warm events is a large increase in the thickness and cross-sectional area of waters between 1.8 and 2.0$^\circ$C shoreward of the DWBC axis.

These changes are summarized in Fig. 12, which plots the transport increments carried in 0.1$^\circ$C potential temperature classes for the composite cold and warm events, as compared to the average DWBC. The total DWBC transport below $\theta = 2.8^\circ$C is similar in each case, between 22 and 25 Sv, as are the relative transport contributions above about 2.2$^\circ$C. However, during the warm events the transport within the 1.8–2.0$^\circ$C NADW core layer is substantially larger (10.2 Sv) than during the cold events (5.8 Sv), while the transport below 1.8$^\circ$C is smaller (2.3 Sv vs 6.4 Sv). These events therefore appear to induce major changes in the density stratification and vertical transport structure of the DWBC.

One possible explanation for these events is that the DWBC is logically composed of two fronts, the deep density front above the DWBC core and the bottom front below the core, that can act more or less independently of one another. The situation during these events appears to be one in which the bottom water front moves onshore and offshore, while the deep front, which builds the deep pressure gradient and therefore controls the location of the DWBC axis, stays nearly stationary. While the details of Figs 11 and 12 are likely to be
Fig. 11. Cross-sections of the inferred DWBC structure based on composite averages of the anomalous "warm" and "cold" events labelled in Fig. 10. Light shading indicates potential temperatures between 1.6 and 2.0°C; heavy shading indicates potential temperatures of less than 1.5°C.

Fig. 12. Volume transport in 0.1°C potential temperature classes for the composite warm and cold events, and for the average DWBC.
incorrect due to the large number of assumptions, we believe them to be reasonably representative of the structural changes that occur during these events. Possible causes for this mode of variability are discussed in the next section.

6. DISCUSSION

These results indicate an equatorward transport of approximately 3 Sv in the DWBC of cold bottom waters below 1.8°C (see also MFJ-92). As shown by WRIGHT and WORTHINGTON (1970), waters colder than 1.8°C occupy two distinct regions in the western basin of the North Atlantic; a northern region north of about 45°N characterized by high salinities, and a southern region south of 30–35°N characterized by low salinities. The low salinity signature of these waters (Fig. 1, see also MFJ-92) clearly identifies them as a mixing product with Antarctic Bottom Water, and precludes a direct connection via the DWBC with the higher salinity North Atlantic Bottom Water. These waters therefore must be recirculating cyclonically in the western basin of the tropical Atlantic. It is unlikely that these waters continue southward across the equator as part of the DWBC, since waters colder than 1.8°C are not found over the 4100 m sill west of the Ceara Rise (see Figs 2a and b), and because previous moored observations over the 4500 m sill east of Ceara Rise showed northward mean flows at all potential temperatures below 1.9°C (WHITEHEAD and WORTHINGTON, 1982). Therefore the Ceara Rise appears to be the southern limit of this recirculation. The northern limit is less clear, although the contours in Figs 2b and c suggest that this deep recirculation extends at least to 14°N, where MFJ-92 found a comparable southward transport in the DWBC of 2 Sv below 1.8°C.

The current meter observations imply large temporal variations in the southward transport of these waters along the western boundary, apparently due to periodic surges of bottom water up onto the continental rise that become entrained in the DWBC. Although these data are too limited to determine the cause of this variability, we can suggest several possible mechanisms.

The first is that these bottom water excursions may be forced by variations in the thickness of the NADW core layer that are either advected along or propagating within the DWBC. In this scenario, a local thickening of the NADW core layer would cause the bottom waters to be squeezed offshore (and vice versa), with the bottom waters responding more or less passively to vertical stretching in the overlying strata.

A second possibility is that these fluctuations could be excited quasi-locally by energetic upper layer eddy processes in the North Brazil Current retroflection region. Here the upper layer currents also are dominated by large fluctuations with 40–60 day periodicity that appear to be related to an instability of the retroflection that causes eddies to be shed at fairly regular intervals (JOHNS et al., 1990). The process is similar to ring formation in the Gulf Stream system, which has been theorized as a primary energy source for the large-amplitude bottom-trapped topographic wave field observed in that region (Hogg, 1981; LOUIS and SMITH, 1982). As in that case, however, this process is difficult to document because the resulting deep motions are in general a random superposition of responses to forcing at different times and places over a broad region. Thus, while there is not much evidence for direct vertical coupling between surface and abyssal layers at the mooring site itself (JOHNS et al., 1990), this does not rule out the possibility that these deep fluctuations could be forced somewhere within this general region.

A third possibility, with larger scale implications, is that these surges of bottom water
are related to variations in the flow of bottom waters over the Ceara sill into the North Atlantic. WHITEHEAD and WORTHINGTON (1982) estimated the mean northward transport of bottom waters (<1.8°C) over the Ceara sill at 4°N to be 1–2 Sv; however, they also observed temporal variations in this transport that were comparable to or larger than the mean. Interestingly, these surges in bottom water flow had a periodicity of about 60 days, similar to the dominant time scale of the deep temperature fluctuations observed here. Using a vertical advection–diffusion model and incorporating a time-varying bottom water flux, WHITEHEAD (1989) predicted that substantial vertical displacements of deep isotherms could occur in the region north of the overflow sill. From a dynamical perspective, one would expect these surges to initially proceed into the Demerara Basin as some form of internal (Kelvin-like) topographic wave along the western flank of the Mid-Atlantic Ridge, eventually spreading into the interior. However, it does not seem reasonable that a large response would be observed north of the sill along the western boundary, particularly at a distance of some 1000 km from the overflow sill.

It should be mentioned that the cause of the 60-day overflow variations at the Ceara Rise remains unknown. WHITEHEAD and WORTHINGTON (1982) speculated that these fluctuations might be driven by a deep (internal) Kelvin wave mode propagating around the Ceara Abyssal Plain south of the overflow sill. An intriguing prospect along these same lines is that the overflow variability at the Ceara Gap could be caused by waves propagating from north of the sill; that is, if the deep fluctuations we observe at 8°N in fact are locally forced, then they could propagate southward along the western boundary to perturb the overflow process at the sill. It is possible that the fluctuations in these two regions are physically unconnected; however, the fact that these available abyssal time series both show the same dominant time scale is difficult to attribute simply to coincidence.

Perhaps the most important issue surrounding these fluctuations is their potential impact on long-term studies of the deep water climatology of the tropical Atlantic basin. While it is not possible to say at this time whether these fluctuations are a localized or basin-wide phenomenon, it is clear that their amplitudes are sufficiently large that they could alias shipboard estimates of the deep water properties around the basin. For example, WHITEHEAD (1989) noted a large inconsistency between the near-bottom temperatures near 10°N measured by the GEOSECS program (BAINBRIDGE, 1981) and by the Transient Tracers in the Ocean (TTO) Tropical Atlantic Survey (WILLIAMS, 1986); the latter survey suggesting that the bottom waters may have retreated equatorward in the intervening decade. Conversely, the STACS measurements, taken some five years after the TTO survey, tend to be more consistent with the GEOSECS data. Are these climate-relevant trends, or just short-term fluctuations? Further progress in understanding the dynamics of these abyssal processes will likely require simultaneous time series measurements at several locations around the tropical Atlantic basin.

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