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Mixing of renewal water flowing into the Caribbean Sea

by Wilton Sturges

ABSTRACT

Observations with moored current meters and STD sections in the Jungfern Passage sill during February 1971 show inflow of $50 \times 10^8$ m$^3$/sec; 5 km downstream, the entrainment of resident water has increased the volume flux by 40%. At times the inflow ceased; the current-meter records suggest a periodicity of ~ two weeks. The inflowing water at the sill is 0.25 to 0.3°C colder than the basin water at sill depth. This signal is eroded to ~ 0.06°C at the downstream section. A clear feature of the observations is that as the flow approaches a geostrophic balance, a bottom boundary layer stretches the flow out on the downslope side, while veering (in the opposite sense) in an upper boundary layer thickens the fluid on the upslope side. To within the accuracy of the observations and the averaging used, the entrainment, as a function of overall Richardson number, is predicted well by the laboratory experiments of Ellison and Turner.

1. Introduction

This paper deals with two related but somewhat different topics. The first is the renewal of deep water in the eastern Caribbean Sea, which has been discussed by Worthington (1966), Sturges (1970), Stalcup, Metcalf, and Johnson (1975), and others. The second topic is the process through which water flowing over a sill (e.g., here in the Caribbean, Mediterranean outflow, etc.) gradually loses its original characteristics by mixing with the surrounding water. There are many situations in which we would like to be able to make dynamically meaningful calculations about mixing. That is, if we wish to know renewal rate, or equivalently, the volume flux of a water mass, we wonder to what extent the inflow must be measured directly. It seems a desirable goal to be able to compute these inflows on the basis of observed large-scale properties, an understanding of the mixing dynamics, and a minimum set of data concerning the inflowing water. There have been some promising laboratory experiments by Ellison and Turner (1959), extended in Turner (1973), and a brief application of some of their ideas in the ocean by James Crease (1965). The present paper describes an experiment designed to test, in a limited way, Ellison and Turner's findings at the much higher Reynolds numbers found in the ocean.

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Figure 1. Location of observations in the Caribbean Sea. Upper portion shows position of Jungfern Passage sill, and the 2000-m isobath. In the lower portion, positions of current-meter moorings (1–12) are shown, with observed mean velocity vectors. Scale is 5' of latitude = 50 cm/sec. Dotted lines show two drifting STD sections. Isobaths are in meters.

2. Current-meter observations

The controlling sill for the Venezuela Basin is at Jungfern Passage, near the Virgin Islands. We made current-meter observations and STD casts from R/V Trident there in January and February of 1971. The deep water comes in from the Atlantic Ocean through the Virgin Islands Basin and flows into the Caribbean Sea through
Wilton Sturges: Mixing of renewal water

1975

Jungfern Passage. Our method was to make STD sections across the inflowing water, and to set out an array of bottom-mounted current-meter moorings (using Geodyne Model 102SE instruments). The first array was set out for two weeks before the STD observations began. Some of the moorings were then recovered and reset, to extend the coverage. Figure 1 shows the location of these moorings. The data are not all simultaneous. Some of the records last for only a few days; some last for several weeks. The times and durations are discussed below (see Table 2; Figure 3). The arrows in Figure 1 show mean velocities for most moorings during the time we made the STD sections. These observations are at ~10 m above the bottom. Most of the current meters are above the bottom logarithmic boundary layer, as discussed below. The bathymetry here is from our own work, with additional information from Stalcup and Metcalf (1973).

When the inflow speeds at mooring number 1 were 20 to 25 cm/sec, the speeds at mooring number 9 were 30 to 35 cm/sec. That is, the flow has accelerated as it has sunk from about 1830 m near the sill to about 2400 m. Mooring number 8 seemed to be on the edge of the flow, showing some bursts of speed ~20 cm/sec and no flow at other times. The position of mooring number 5 is correct as shown in the figure, but the interpretation of its flow direction shown in Figure 1 requires explanation. The inflow direction at mooring number 1 is shown for the time of the STD work. For the two weeks earlier, during the time mooring number 5 was in place, the flow direction at mooring number 1 was ~30° more toward the south. That is, the inflow at the sill was directed more nearly toward mooring number 5.
Figure 3. Inflow-directed component of the velocities observed at various moorings during the latter part of the STD work. Times of the STD stations are shown at top.

than it was during the times of observations at moorings number 8–10, and the STD sections. During this later time, the flow shown in Figure 1 for mooring number 5 would probably be found roughly a mile and a half to the west of the position shown.

Figure 2 shows the speeds obtained at mooring number 1 resolved into components about the main inflow direction. The peak speeds are near 30 cm/sec; the average is 21.5 cm/sec. During the last week of this record, while STD observations were being made, the flow speed was fairly constant. This same indication of fairly steady flow was seen at other current meters while we were making drifting STD observations, as shown in Figure 3.

There is some indication of a periodicity of about two weeks. The first day and a half of record 1 indicate no inflow; the flow begins shortly after the time of spring tides, and the minimum inflow coincides with the neaps. The record is too short to permit any definite conclusion, but it is very suggestive that Stalcup, Metcalf, and Johnson (1975) observed similar variability. The spectra from the current-meter records show a pronounced peak at a period of 6.2 hours, similar to that reported by Stalcup et al., which seems to be a large-scale seiching motion.

The oscillations near the middle of this record (1) are semidiurnal with amplitudes ~ 8 cm/sec. Moorings number 7 and 9 were launched 45 hours after these oscillations stopped. They show a mean flow of 20 to 25 cm/sec with superimposed near-inertial oscillations of ~ 10 cm/sec.

There were temperature recorders on most of the current meters also. The change in temperature from a period of no inflow to a period of maximum inflow was 0.32°C at mooring number 1, although a typical value during the time of the observations was approximately 0.25°C colder than background. During the large semidiurnal excursions in velocity, the temperature records showed the same features. This result agrees with the findings of Stalcup, Metcalf, and Johnson (1975). This finding also supports an assumption made earlier (Sturges, 1970) that the sill would exert a rectifying effect on the inflow.

At current meter 5 there were observations at 10 m and 50 m above the bottom. At the lower current meter, during the steadiest flow, speeds were 20 cm/sec; at
the upper one the speeds were only 5 cm/sec. The direction at 10 m was \( \sim 40^\circ \) to the left of the flow direction at 50 m.

3. STD observations

Figure 1 shows the position of two STD sections across the inflow. These sections were made by allowing the ship to drift across the flow (the wind was blowing in the right direction) and by lowering the STD repeatedly through the cold layer about every 10 to 15 minutes. In addition to the normal recording of the data, frequency counters were used to monitor the \( f-m \) signals from the STD. At the bottom of each brief lowering, the frequency from the pressure signal was recorded, and this was used to determine the depth along each STD section. The speed of drift was typically 0.7 to 1 knot.

The inflowing water is characterized by a strong temperature signal of \( \sim 0.25^\circ \text{C} \) and a salinity difference of \( \sim 0.02^\circ/\text{oo} \). The cold inflowing water was observed to have a consistent minimum temperature near the sill of \( \sim 3.80^\circ \text{C} \), and occasionally a few hundredths cooler. The potential temperature is 3.63\(^\circ \text{C} \), salinity \( \approx 34.985^\circ/\text{oo} \). The temperatures are in accord with those of Stalcup, Metcalf, and Johnson (1975) who report minimum potential temperatures of 3.62 to 3.67\(^\circ \text{C} \). The inflowing water is colder and more saline than the water at the sill, and also colder by \( \sim 0.2^\circ \text{C} \) than the deep water in the eastern Caribbean, and more saline by \( \sim .005^\circ/\text{oo} \) (see, for example, Sturges 1965, Figure 4; 1970, Figure 1; Worthington 1966, Figures 3, 5). The density difference at the sill is \( \sim 2.5 \times 10^{-5} \text{ gm/ml} \) from the temperature difference, plus \( \sim 1.5 \times 10^{-5} \text{ gm/ml} \) from the salinity difference. To the best of our ability to resolve salinity in the observations reported here, the temperature-salinity correlation did not change within the inflowing water from one side of an STD section to the other, nor from one section to the next. The temperature sections, therefore, convey all the STD information.

The two sections presented here are the one nearest the sill and the one farthest downstream, and are the most complete ones we made. Figure 4 shows the tempera-
Table 1. Background temperature from STD observations beyond the cold inflow water, relative to which the temperature deficits are computed. These values are thought to be accurate to within ± 0.01°C.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Temperature (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1550</td>
<td>4.20</td>
</tr>
<tr>
<td>1600</td>
<td>4.16</td>
</tr>
<tr>
<td>1700</td>
<td>4.12</td>
</tr>
<tr>
<td>1800</td>
<td>4.09</td>
</tr>
<tr>
<td>1900</td>
<td>4.08</td>
</tr>
<tr>
<td>2000</td>
<td>4.08</td>
</tr>
<tr>
<td>2200</td>
<td>4.08</td>
</tr>
<tr>
<td>2400</td>
<td>4.08</td>
</tr>
</tbody>
</table>

ture section nearest the sill, presented as the temperature deficit between the inflowing water and the normal Caribbean water, near the sill, at the same depth outside this inflowing layer. The mean values of background temperature, shown in Table 1, are mainly based on data from the drifting stations, taken before reaching the cold inflowing layer. Although there is internal wave noise in Figure 4, the basic signal is a layer of cold water with its temperature deficit decreasing upward.

This section (shown in Figure 4) was divided (somewhat arbitrarily) horizontally into thirds to average the temperature, and these average values have a nearly exponential distribution, as discussed below. There is very little slope of the isotherms; the Rossby number, $U/fL$, is estimated to be about 1.5 to 2. The flow probably behaves as a jet squirting over the sill. For a two-dimensional jet, a similarity solution suggests that the velocity distribution in the upper layer should be described by a hyperbolic secant. For our purposes there is a negligible difference between that and the exponential. In Ellison and Turner's laboratory results, they found that the upper limit of the density difference coincided fairly well with the upper level of the velocity profile also (see their Figure 4).

We expect the Rossby number to be fairly large in the region of the sill, as mentioned above; we estimate 1.5 to 2 as a lower limit. However, the dominant period of motion that was observed at current meter 2 (near the eastern edge of the sill), was 18 hours, or approximately one-half the local inertial period. This observation suggests that the local Rossby number was very nearly 1. So far as I am aware, there is no theory which would predict an upper limit of 1 on the Rossby number in this situation.

Figure 5 shows the temperature distribution observed ∼ 3½ miles downstream. The flow is out of the page. We see that several features have changed, compared with the section across the sill. The section is wider, although the average height has not changed. The average temperature difference has been eroded away. There is a fairly thin layer near the bottom where the deficit is ∼ 0.08°C, although there are isolated regions where it reaches 0.1°C. An obvious feature is that on the lower side, the cold layer has been stretched out. This effect is probably caused by the bottom
boundary layer. It also can be seen that the upper side (to the left, in the figure) appears to be about 3 times as thick as at the upstream section. This effect could be caused by veering in an upper boundary layer or by entrainment. These effects are discussed in the following sections.

4. Bottom boundary layer

There are two boundary layers in this inflow. The bottom boundary layer is of the kind that has been studied by the meteorologists. Recently Weatherly (1972) has studied the bottom boundary layer under the Gulf Stream; the interpretation of the observations reported here depends heavily on his results.

The shear, absolute speeds, and vertical stability in the Caribbean inflow situation observed here are remarkably similar to those observed by Weatherly. It seems to be a safe assumption that the ratio \( u_\star / V_g \) would also be essentially the same: \( u_\star \) is the "friction velocity" appropriate to the bottom stress \( (\tau_0/\rho)^{1/2} \) and \( V_g \) is the geostrophic velocity in the interior. The ratio \( u_\star / V_g \) was taken to be 0.035, the value determined by Weatherly. The thickness of the logarithmic boundary layer is then computed as

\[
\delta_L = 2u_\star^2 / fV_g.
\]

For (an assumed) \( V_g = 20 \text{ cm/sec} \), \( \delta_L = 12 \text{ m} \), so at several of these moorings the observations were made barely within the log layer. Assuming that the velocity distribution is logarithmic, if the velocity is known at some height, such as 10 m, then \( V_g \) can be computed; these are given in Table 2. Even in the extreme case, however, at mooring number 9, the thickness of the log layer is 16 m, and the dif-
Table 2. Mean speeds observed at the current meters. Calculations of log layer and free stream speed \( V_g \) are based on bottom-stress data of Weatherly (1972); see text. The column headed Date gives day, month and hour of the start of useful data; then the number of days of data before recovery.

<table>
<thead>
<tr>
<th>Mooring No.</th>
<th>Date</th>
<th>Height Above Bottom (m)</th>
<th>Observed Speed (cm/s)</th>
<th>Thickness of Log Boundary Layer (m)</th>
<th>Computed ( V_g ) (cm/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>22 I 1500/23</td>
<td>9</td>
<td>21.5</td>
<td>13.</td>
<td>24.</td>
</tr>
<tr>
<td>2</td>
<td>22 I 1700/16.2</td>
<td>9</td>
<td>8.9</td>
<td>5.5</td>
<td>8.9</td>
</tr>
<tr>
<td>3</td>
<td>22 I 2300/12.7</td>
<td>10</td>
<td>20.</td>
<td>12.</td>
<td>22.</td>
</tr>
<tr>
<td>4</td>
<td>22 I 2300/12.7</td>
<td>50</td>
<td>6.5</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>5</td>
<td>3 II 2100/6.4</td>
<td>10</td>
<td>20.2</td>
<td>12.</td>
<td>22.</td>
</tr>
<tr>
<td>6</td>
<td>8 II 2200/3.5</td>
<td>10</td>
<td>21.2</td>
<td>13.</td>
<td>23.</td>
</tr>
<tr>
<td>7</td>
<td>12 II 1600/3.7</td>
<td>9</td>
<td>26.5</td>
<td>16.</td>
<td>29.4</td>
</tr>
<tr>
<td>8</td>
<td>14 II 0100/3.5</td>
<td>10</td>
<td>12.7</td>
<td>8.</td>
<td>12.7</td>
</tr>
<tr>
<td>9</td>
<td>16 II 0100/1.6</td>
<td>10</td>
<td>15.</td>
<td>9.</td>
<td>15.</td>
</tr>
</tbody>
</table>

The difference in speeds is only 3 cm/sec between \( V_g \) and the speed observed 10 m above the bottom.

The total thickness \( H \) of the bottom boundary layer is

\[
H \sim k u_s / f
\]  

(2)

for an observed \( V_{10} = 20 \text{ cm/sec} \), hence \( V_g = 22 \text{ cm/sec} \) and \( H \sim 70 \text{ m} \). It seems correct to interpret the flow to the left of the bottom contours, at the deeper section, as the result of bottom friction.

At mooring number 5, the veering angle between the current meter at 10 m and at 50 m is highly variable; during the periods of nearly steady flow the angle was \( \sim 40^\circ \). The direction of the isobaths changes along the flow, but at moorings number 7 and 9, where the bottom direction is changing less rapidly, the flow (near the top of the logarithmic boundary layer) lies in a direction \( \sim 20^\circ \) to the left of the isobaths. One might speculate, therefore, that of the 40° veering observed at mooring number 5, about half results from veering in the bottom boundary layer; and that roughly half occurs in the upper boundary layer.

5. Upper boundary layer

In the initial stages of this flow, as in the flow over all sills, there is some sinking of the water parcels, as well as mixing. If we consider the central part of the flow—that is, away from the lateral edges, the following terms, in the downstream momentum equation seem important,

\[
\frac{\nu \partial v}{\partial y} + \frac{w \partial v}{\partial z} + \frac{\partial (w'v')}{\partial z} + fu = -\frac{1}{\rho} \frac{\partial p}{\partial y}
\]  

(3)
where the main inflow is in the y-direction; z is positive upward, in a righthanded coordinate system. The pressure-gradient term will be a "driving" mechanism as the dense water initially sinks. By estimating the average density deficit in the central portion between the two STD sections (.86 x 10^-5 gm/ml), and knowing the slope of the isotherms along the flow, the maximum value of this term is found to be \( \approx 9 \times 10^{-4} \) cm/sec^2. An average value, higher up in the flow, would be \( \approx 3 \times 10^{-4} \) cm/sec^2. The first term on the left-hand side of (3) can be found by comparing the speeds at current meters 1 and 9, to obtain \( 2.6 \times 10^{-4} \) cm/sec^2. By using mooring number 10, where the flow has not sunk deeper, this term would be negative. The second term on the left represents a loss of momentum by the entrainment of ambient fluid. From the calculated total entrainment flux, I estimate \( w \) at the top of the layer to be \(-.06 \) cm/sec; using half that value for a mean, and taking the layer to be 100 m thick, this term is \( 0.8 \times 10^{-4} \) cm/sec^2, or fully one-third as large as the horizontal acceleration term.

The third term in (3) is the eddy stress term that, physically, is equivalent to the \( A_z \frac{\partial^2 v}{\partial z^2} \) term in the classical wind-driven surface Ekman layer. As long as there is turbulent mixing across the upper boundary of the inflowing water, we should expect this term to contribute to the formation of an upper boundary layer in which there is turning of the flow direction. The magnitude of this term cannot be reliably estimated from the data in hand. Everywhere above the level where its maximum value occurs, the sign of this term will be negative, contributing in the classical way to a positive value of the Coriolis term.

As a result of these arguments, there are found to be three terms in (3) which may contribute toward the production of a positive value of the Coriolis term. In the deep central portion, there appears to be an excess of the pressure-gradient force term over that required to accelerate the flow. On the up-slope side of the flow, where there is a much reduced horizontal pressure gradient, the horizontal acceleration term, now negative, has the appropriate sign. And in all cases, above the maximum value of \( w' v' \), the vertical eddy stress term contributes in the correct sense.

It seems a reliable conclusion, therefore, that the Coriolis term, \( f u \), has the sign that requires cross-stream flow in the up-slope direction, opposite that in the bottom boundary layer. The effect of this upper cross-flow would be the accumulation of the less dense inflowing water on the up-slope side of the inflowing layer (to the left-hand side in Figure 5) as observed. Because there are substantial uncertainties in the numerical results, a veering angle of 20° as inferred at mooring number 5 is compatible with these calculations, but so is 75°. The essential point here is that there is cross-stream flow in the up-slope direction; there is an upper boundary layer in which the sense of rotation of the velocity vectors is opposite to that in the lower. The magnitude of this term is \( v \sin \theta f \); for \( v = 14 \) cm/sec, \( \theta = 20° \), we obtain \( 2.1 \times 10^{-4} \) cm/sec^2.
Figure 6. Average temperature distribution at section 32, downstream, open figures; and at section 39, in the sill, filled figures. The circles are for the eastern third of each section; triangles, middle third; and squares, western third. A decay of $e^{-nZ}$ is shown for comparison. The values of $T_0$, $Z_0$, and $D$ are given in Table 3.

As the flow begins to approach a geostrophic balance, the momentum balance in the cross-stream direction will be expected to be much more nearly geostrophic than in the downstream direction. As a result, it is consistent to estimate the vertical shear, in the downstream direction, by the usual thermal wind equation, and simultaneously to expect cross-stream flow caused by the veering.

The average height of the isotherms above the bottom can be used to determine a scale height for the temperature distribution of the form:

$$\Delta T = \Delta T_0 \exp - \frac{\pi (Z - Z_0)}{D} \tag{4}$$

The temperature has been averaged vertically, by dividing the section into thirds, and the results are shown in Figure 6 and Table 3. It is clear from visual inspection of STD sections 39 and 32 that the downstream one is greatly distorted compared to the upstream section. Table 3 shows that the average scale height changes only slightly. The temperature deficit, however, is decreased $\sim 75\%$ on the western side, and $\sim 60\%$ on the eastern side.

Using the scales given in Table 3 for the middle of the downstream section, an exponential velocity distribution would reduce the 20 cm/sec observed at 10 m, on

<table>
<thead>
<tr>
<th>STD 39</th>
<th>D</th>
<th>M</th>
<th>East</th>
</tr>
</thead>
<tbody>
<tr>
<td>at sill</td>
<td>70 m</td>
<td>130</td>
<td>146</td>
</tr>
<tr>
<td></td>
<td>14 m</td>
<td>28</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>.2°C</td>
<td>.23</td>
<td>.15</td>
</tr>
<tr>
<td></td>
<td>.188 C</td>
<td>.257</td>
<td>.175</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>STD 32</th>
<th>D</th>
<th>M</th>
<th>East</th>
</tr>
</thead>
<tbody>
<tr>
<td>downstream</td>
<td>198 m</td>
<td>68</td>
<td>37</td>
</tr>
<tr>
<td></td>
<td>7 m</td>
<td>16</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>.04 C</td>
<td>.06</td>
<td>.06</td>
</tr>
<tr>
<td></td>
<td>.043 C</td>
<td>.09</td>
<td>.068</td>
</tr>
</tbody>
</table>

Table 3. Average scale values for temperature distribution in the two STD sections, from a fit to equation (4). The quantity $\Delta T_z = 0$ is the average temperature deficit, for that part of the section, at the bottom. The instrument was usually about 1 to 2 m from the bottom.
mooring number 5, to 4 cm/sec at 50 cm. This result compares favorably with the 4 to 5 cm/sec observed, during a time of nearly steady flow.

6. Geostrophic adjustment

In Figure 5 it is clear that there is substantial slope to the isotherms. The vertical geostrophic shear can be calculated, and (assuming the fluid outside to be at rest) compared with the observed speeds. At STD 32, this calculation gives 26 cm/sec in the central region, and 33 cm/sec along the eastern side. The computed speeds in the east are a few cm/sec higher than at mooring number 9. The acceleration term has the magnitude to account for the discrepancy. That is, the Rossby number could be 0.2 to 0.4; because STD 32 was made while none of these moorings were in place, however, it seems pointless to push the comparison further.

7. Entrainment and total mass inflow

The inflow volume flux can be calculated using the observed velocities and assuming an exponential vertical shear which is consistent with the temperature distribution. The inflow thus determined, at the sill, is $49 \times 10^3$ m$^3$/sec. At the last section farther downstream, this inflow has increased to $70 \times 10^3$ m$^3$/sec. Both values are uncertain to at least 100%. In other words, the flow has entrained about 40% of its original amount.

One question about the comparison of these two sections is the steadiness of the inflow. Figure 3 shows a composite of the inflow speeds at the four moorings that were in place during the time of these STD observations. The downstream section (STD 32), was made during an interval of slightly reduced inflow. (A lag of about 7 hours from the sill to that position is appropriate.) Because the reduction in inflow at that time is barely evident at mooring number 1, it may be that the flow axis had meandered slightly to the west. Nevertheless, it seems safe to assume that the inflow during STD 39 (at the sill) was certainly as strong as for STD 32. Therefore, the observed increase in transport between the two sections is a lower bound on the entrainment; the increase is not merely time variability.

Ellison and Turner (1959) found that the entrainment across a stratified interface is a function of the overall Richardson number of the flow, which they defined as (their equation 5)

$$\frac{\Delta h \cos \alpha}{V^2} = Ri$$

where $\Delta$ is an average density deficit across the layer of scale thickness $h$ and a scale velocity $V$; $\alpha$ is the bottom slope. They observed in their experiments that $V$ was generally $\sim 0.7$ times the maximum velocity. By integrating the (assumed) exponential velocity profile above a uniform lower layer of thickness $Z_0$, the average $V$ in our results (by their definitions) is found to be $0.67 \pm 0.07$ times $V_g$, in surprisingly good agreement with their results.
Table 4 gives the calculated transports within each third of the two sections, and the Richardson numbers computed from Ellison and Turner's definitions. The transport increase is not uniformly distributed across the section, but is concentrated in the western side. This result may be partly the result of observational uncertainty. However, the boundary-layer veering will tend to force the entrained water toward the western side of the section. Furthermore, the lower boundary layer will drain fluid away toward the downhill, eastern side. The observations are consistent with these ideas.

The entrainment coefficient $E$ was defined (their equation 7) as

$$\frac{1}{V} \frac{\partial (Vh)}{\partial x} = E$$

where (their notation) $x$ is the downstream coordinate. They treated a two-dimensional flow. To take changing width into account, one may multiply the downstream value of $Vh$ by the ratio of the width of the two sections. Using the mean value $V = 12.4$ cm/sec for the six pieces of the sections, we find a mean $E = 0.005$. The mean $Ri$ is 0.5, or 0.6 if the single low value 0.05 be excluded. These mean values fall surprisingly well on the curve of Ellison and Turner's Figure 10. It should be clear that this crude averaging of the results may not be what one would prefer, but a more precise comparison asks too much of the data. The essential result of this comparison, then, is that to within the accuracy of our observations, the entrainment coefficients of Ellison and Turner also apply to the much larger Reynolds-number flows in the ocean.

It might have been assumed, a priori, that the effects of rotation would be dominant in the large-scale aspects of the flow, but would be less important in the small-scale features that control the turbulent mixing. This idea seems to be borne out. However, the modification of the overall flow—in the two rotationally-modified boundary layers—does affect the Richardson number.

P. Smith (1974) has also found order-of-magnitude support for Ellison and Turner's entrainment coefficients in an analysis of the Mediterranean and North Atlantic overflows. It is to be hoped that sufficient evidence can be accumulated to allow further and more accurate tests of this concept, as it seems to be a most promising method for parameterizing—in a clear, precise, and predictable manner—the effects of turbulent mixing.
8. Renewal rate

The volume flux of renewal water found here is approximately half the required inflow estimated earlier (Sturges, 1970) on the basis of heat-budget and oxygen-budget considerations. I do not believe the two values are significantly different, however, so that a renewal time of 500 years, plus or minus a factor of 2, is consistent with the available data.

The area between the 2000 m and 3000 m isobaths in the Columbia and Venezuela Basins is $\sim 30 \times 10^4$ km², for a total volume of $\sim 30 \times 10^4$ km³ between those two isobaths. This volume, divided by the flux at the sill, gives an estimated renewal rate, for the thousand meters below sill depth, of only $\sim 200$ years.

The mixing requires a change, however, in our rather simple heat budget calculations. The original cold, dense water is cooler by $\sim 0.25°C$ than the deep water. As it entrains warmer water, this (mixed) warmer water is carried down also, so that the "deep water" thus formed has a substantially reduced cooling capacity. For the $\sim 20 \times 10^3$ m³/sec of water that is entrained, between the sill and the downstream section, one wonders: what was its source?

The renewal rate of deep water appears, therefore, to be a strong function of depth. A large part of the final mixture of deep water will be water from below sill depth, which has been entrained and carried down. Until we can quantify the mixing dynamics better, it will be difficult to make meaningful calculations of the budgets of heat or mass in deep basins.

9. Causes of inflow variability

The inflow observed during this experiment certainly appears to be variable. Stalcup, Metcalf, and Johnson (D75) also observed sporadic inflow. One source of this variability may be the spring-neap tidal variation. That is, very strong tides may initiate a flow which, when begun, continues until some limiting factor sets in; e.g., the drainage of cold water from the Virgin Islands Basin changes the driving pressure head.

The slope of the sea surface across the Caribbean Sea is about 25 to 30 cm, as a result of the prevailing surface currents (see Gordon, 1967). It seems likely that periods of strong surface flow—which would be caused by a period of strong trade winds—could lead to a larger baroclinic pressure head on the northern side of the Caribbean. If this pressure gradient extends to 1800 m, it might lead to periods of reduced inflow, and conversely. If the trades blow less strongly toward the Caribbean during the fall, one might speculate that the greatest inflow of deep water would be in that season.

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