Surges of Antarctic Bottom Water into the North Atlantic*

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11 May 1987 and 23 January 1989

ABSTRACT

Current meter records show that Antarctic Bottom Water surges into the western North Atlantic with roughly a sixty-day period. A time-dependent mass budget which incorporates estimated volume fluxes from geostrophic calculations, surges with a sixty-day period, a known time-average volume of the very coldest water in the North Atlantic, and a constant mixing coefficient predicts vertical excursions of 160–230 meters for the 1.1° isotherm and 45–60 meters for the 1.2° isotherm. Available data do not reveal such an excursion. The reason for the lack of the excursions, is that earlier estimates of volume of the very coldest water were too small. Corrected tables are presented. The disagreement between the current meter results and the geostrophic calculations remains.

1. Introduction

Bottom water ($\theta < 1.9°C$) in the western North Atlantic has partial Antarctic origin and flows northward over the Ceara Abyssal Plain into the North Atlantic through a bathymetric saddle point, roughly centered at 4°N 39°W. Estimates of volume flux of this water by Whitehead and Worthington (1982) are 0.8×10^6 $m^3 s^{-1}$ based upon a 360-day current meter dataset and 1.98×10^6 $m^3 s^{-1}$ based upon geostrophic estimates. The disagreement arose almost entirely from an estimated flux of approximately 1×10^6 $m^3 s^{-1}$ for water colder than 1.2°C in the geostrophic calculations. This estimate was made because water warmer then 1.2°C had a substantial northward flow, the vertical shear from thermal wind for water colder than 1.2°C was close to zero, and the cross-sectional area was quite large. On the other hand, the deepest current meter (near $\theta = 1.055°C$) showed little time-averaged northward flow. The low value of the mean northward component of the velocity might arise if the current meter was in the bottom boundary layer. However, this possibility is eliminated because the surge amplitudes (i.e., the rms signal) were as large as those in warmer water above.

Since the geostrophic and current meter estimates differed so greatly for water colder than $\theta = 1.2°C$, eddy mixing coefficients which were deduced from the two datasets differ by factors of between 3 and 35. These values of eddy mixing coefficients have been taken by Gargett (1984) and Gregg (1987) to be representative of the deep ocean, since direct measurements have mostly been done above 2000 m. McDougall and Whitehead (1984) examined the implication of the two estimates in conjunction with what is presently known about mixing coefficients from laboratory studies, and concluded that the geostrophic numbers yielded more "reasonable" eddy coefficients. Unfortunately, the argument is not conclusive, and the disagreement of between 3 and 35 in the eddy mixing coefficients still hold. This present study is the result of an attempt to generate a more precise estimate.

In order to investigate the implications of these transport estimates to conditions downstream of the sill, a calculation of the time response of the volume of the deepest water mass to an oscillating source is given in section 2. These calculations were motivated by the fact that all current meters in Whitehead and Worthington showed sizeable oscillations with approximately a sixty-day period. In comparison, the residence time found by dividing the volume of the water mass by the geostrophic estimate of volume flux for water colder than 1.1° was 44 days. Section 3 shows that a sixty-day periodic surge into a basin where the water only has a 44-day residence time would probably cause a large fluctuation of its volume which should be seen as a raising and lowering of isotherms. Histor-

* This note had been intended for the volume of the *Journal of Physical Oceanography* dedicated to the memory of Adrian Gill, a wonderful friend who inspired me with his brightness combined with a quite wit, an endearing modesty, and personal humility. This is an attempt to find actual overlap between prediction based on a simple model and factual ocean data. This is always a challenge, and one which Adrian would have enjoyed. I am glad to have known Adrian Gill, and I do miss him.

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ical observations of the depths of the 1.1 and 1.2 isotherms are reviewed in section 4. They did not reveal any evidence for the large volume fluctuation predicted in section 3. Although it was first thought that these considerations would convincingly rule out the geostrophic estimates, a closer look at the data, as discussed in section 5, reveals that there were large errors by Whitehead and Worthington in the estimates of the volumes of the water masses colder than 1.2°C. These errors were due to the sparsity of the data available at that time. New estimates of the volume are made and when their values are used in the model, the predicted magnitude of the vertical excursion of isotherms is too small to detect. Although the conflict between current meter and geostrophic estimates remains, the new volumes lead to estimates of smaller (by factors of 2 to 4) mixing coefficients in section 6. All these values, whether for current meters or geostrophic estimates, are considerably lower than the values calculated, in a similar way, by Hogg et al. (1982) for flow in the Vema channel going into the Brazil Basin, and by Saunders (1987) going northward in Discovery Gap into the eastern North Atlantic.

2. The calculation

In this section we construct a simple model in which an oscillating source introduces cold deep water into a basin whose steady state is maintained by downward diffusion of heat. Consider first the relationship between depth of the water mass and volume flux for a steady problem. Take as a simple model a box of water (Fig. 1c) insulated on the sides and with uniformly upwelling water of velocity W and bottom temperature θ_1. Let the water exit at the top with temperature θ_2. Take as a model a basin into which bottom water is flowing through a shallow channel at flow rate U then plunging down to the bottom. The deep water thus supplied is then upwelling with velocity W and passing through isotherms; the upward advection of cold water is balanced by downward (turbulent) diffusion of heat. During some periods, $U = U_1$ is fast (Fig. 1a), so $w = w_1$ is large, hence volume V_1 of the coldest deep water is large. At other times (Fig. 1b), $U = U_2$ is slow; so w_2 is small and V_2 is different. We first show that for an advection–diffusion model, $V_2 < V_1$. The temperature distribution is determined by solving the differential equation for temperature θ

$$W\frac{d\theta}{dz} = \kappa\frac{d^2\theta}{dz^2},$$

subject to the boundary conditions $\theta = \theta_1$ at $z = 0$, $\theta = \theta_2$ at $z = L$, where κ is thermal diffusivity. The solution is

$$\theta = \{[\theta_1 - \theta_2]\exp(Wz/\kappa) - \theta_1\exp(WL/\kappa) + \theta_2\}/[1 - \exp(WL/\kappa)].$$

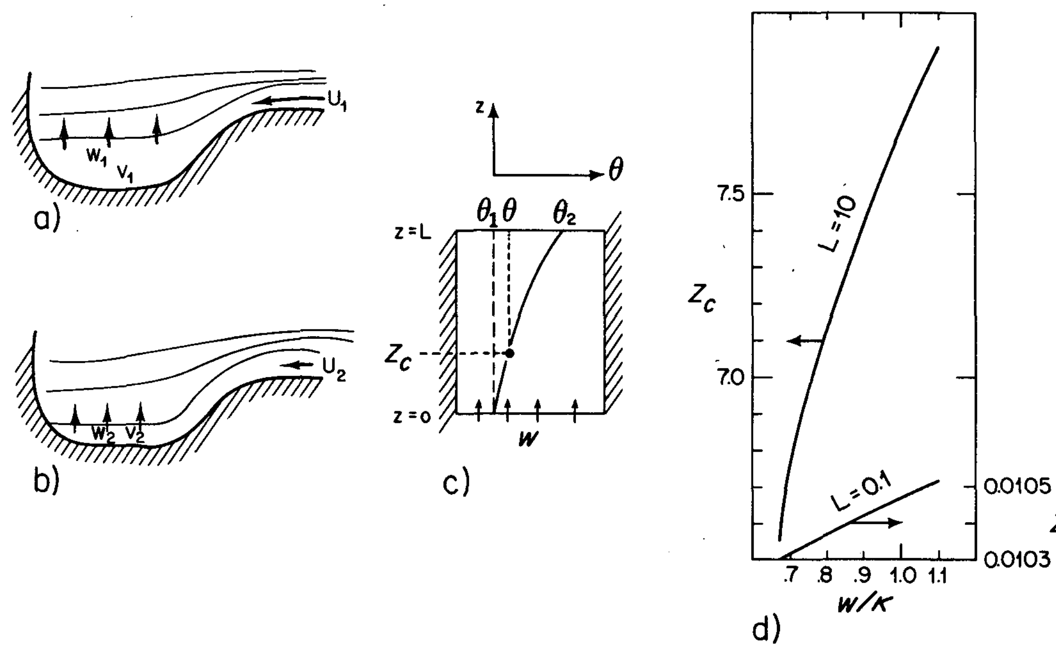

FIG. 1. Sketch of the model. (a) Flow of cold water into a basin with large velocity U_1. It produces large upwelling velocity W_1, which makes volume V_1 of the coldest water large. (b) For slow flow U_2, W_2 and V_2 are smaller. (c) A side-insulated box which supports the pictures in (a) and (b) has uniform upwelling with water coming in with $\theta_1 = 0$ at $z = 0$ and leaving with θ_2 at $z = L$. (d) Solutions to the equation for the height z_c of the $\theta_1 + 0.1\,(\theta_2 - \theta_1)$ isotherm as a function of upwelling velocity for two values of L. Height z_c monotonically increases with W/κ.

We ask, what is the depth of a fixed isotherm as a function of W? Let us choose as our selected isotherm the height z_c at which $\theta_c = \theta_1 + 0.1\ (\theta_2 - \theta_1)$. The solution for z_c is

$$z_c = [\kappa/W] \ln\{1 - (0.1)(1 - \exp(LW/\kappa)\}.$$

For a given value of L, z_c monotonically increases with W and decreases with κ. To show this, one can expand the above equation in a Taylor series, or one can directly solve this with trial values and a hand calculator. Two such results are shown in Fig. 1d. The inverse length scale W/κ is varied through a parameter range close to one (for convenience) and the monotonic nature is clear.

Turn now to a second time-dependent model in which the rate of change of volume of water colder than a given temperature is equal to the difference between the volume flux of the source $Q_s(t)$ and the volume flux rising through a given isotherm $Q_r(t)$

$$A dh/dt = Q_s(t) - Q_r(t).$$

where A is the cross-sectional area of the basin.

For $Q_r(t)$, use the results of the steady-state model solved above by setting the depth of water $h(t)$ linearly proportional to the instantaneous value of a time-dependent upwelling velocity $W_r(t)$. Thus

$$Q_r(t) = W_r(t)A = Ch(t).$$

One can think of C as an eddy or mixing diffusivity integrated over area A.

For volume flux of the source, we will combine a steady source with a simple sinusoidal signal $Q_s(t) = Q_1 + Q_2 \sin(\omega t)$.

These considerations result in a simple differential equation describing the height $h(t)$

$$A \frac{dh}{dt} = -Ch + Q_1 + Q_2 \sin\omega t \tag{1}$$

where $\omega = 2\pi/P$ is the frequency and P is the period of the fluctuating source. We will take in general the case where A and C are some integer power of h

$$A = \alpha_n h^n$$

$$C = \chi_n h^n.$$

The exponent represents different bathymetric configurations for the downstream basin. With a flat bottom, $n = 0$. For $n = 1$, there is a V-shaped trough with vertical end walls. For $n = 2$, the basin is either a conical depression or a parabolic trough with vertical end walls. Equation (1) is written as

$$\frac{\alpha_n}{(n+1)} \frac{dh^{n+1}}{dt} = -\chi_n h^{n+1} + Q_1 + Q_2 \sin\omega t. \tag{2}$$

The solution to (2) has a steady term

$$H^{n+1} = Q_1/\chi_n, \tag{3}$$

and a time-dependent term

$$(h')^{n+1} = \frac{Q_2 \sin(\omega t + p)}{\omega \dfrac{\alpha_n}{n+1} \sin p - \chi_n \cos p} \tag{4}$$

where $\tan p = \omega\alpha/\chi_n$.

It is useful to work with

$$T = \frac{\alpha_n}{\chi_n} = \frac{\alpha_n H^{n+1}}{Q_1} \tag{5}$$

from (3), which is a residence time for the steady component flow. Substituting (5) into (4), the ratio of the fluctuating to the mean depth of an isotherm is found as

$$\left(\frac{h'}{H}\right)^{n+1} = Q_2 \sin(\omega t + p) \Big/ \left[Q_1\left(\frac{\omega T}{n+1} \sin p - \cos p\right)\right]. \tag{6}$$

In general, if current meters have measured Q_2, Q_1 and ω, and if hydrographic data can be used to deduce h', h and A, Eq. (6) can be tested.

3. Test using existing estimates

Data to estimate T, ω, and the ratio Q_2/Q_1 can be found in Whitehead and Worthington. We will use these to predict h' for two cases, one using their current meter estimates of volume flux as a basis for measuring T (Table 1a) and the other using geostrophic estimates (Table 1b). We will do this in detail for $n = 0$, and then discuss other values of n. To estimate residence time T, column 2 in both tables gives estimates of the volume of water north of 4° colder than a given temperature. This was found by cumulatively adding from below numbers for volume estimates between isotherms from column 15 in Table 3 of Whitehead and Worthington. Column 3 lists the estimate of volume flux ($m^3\ day^{-1}$) of water being advected upward across each isotherm as determined from an estimate of flow into the North Atlantic in column 4 of Table 3 of Whitehead and Worthington. Residence time T in column 4 is found by dividing the numbers in column 2 with corresponding ones in column 3. (Note that this definition of residence time is based on total volume below an isotherm, rather than volume between isotherms. Therefore, our residence time values differ from those in Table 3, column 18, of Whitehead and Worthington.)

The remaining two columns in Table 1 give the remaining parameters needed to evaluate h'/H in Eq. (6). The parameter combination $\omega T = 2\pi T/P$ is given in column 5. A period of $P = 60$ days for the oscillating source is obvious from Fig. 2. This value was used in column 5 for all cases, since all other current meters displayed the same periodic surges. To get the ratios

TABLE 1. Some parameters and the predicted vertical excursions of isotherms $2h'$ for water colder than 1.3°C in the North Atlantic using data in Whitehead and Worthington.

1 θ (°C)	2 Volume (10^{12} m^3)	3 Volume flux upward (10^{12} m^3 day^{-1})	4 T (days)	5 ωT	6 Q_2/Q_1	7 $2h'$ (m)
		(a) Current meter results				
1.3	60.2	0.020	3010	315	1.63	1.4
1.2	11.2	0.010	1120	117	2.2	2.2
1.1	2.2	0.003	733	77	4.5	9.3
		(b) Geostrophic results				
1.3	60.2	0.109	552	58	1.63	13
1.2	11.2	0.096	117	12	2.2	44
1.1	2.2	0.050	44	4.6	4.5	158

of fluctuating to steady flow shown in column 6, values of velocity were used directly from current meter data. The steady component is estimated for each tenth of a degree interval in Table 2 of Whitehead and Worthington. For the time-dependent component, the measured peak to trough amplitude for the northward component of the time dependent current divided by two was taken from Fig. 2 to be 8.5 cm s^{-1}. The same values of the ratio Q_2/Q_1 were used for both the current meter and geostrophic estimates, since only current meters have estimated Q_2.

At first glance, the use of the measurement Q_2/Q_1 for the geostrophic calculation looks suspicious, as we are dividing by the current meter value of Q_1 which is small. However, this is not so drastic as it first appears. We must adopt some value for the ratio of the fluctuating to steady flows and the only information available is from the current meters. Thus, we have explicitly arranged it so that Q_2 is divided by Q_1, so that this measurement can be incorporated. Note however that another Q_1 is in the denominator of the expression for T. The value of this other Q_1 will differ depending upon whether current meter or geostrophic results are used. Assume that Eq. (6) uses some explicit estimate of the fluctuating volume flux. The current meter results just use $Q_{2\text{current meters}}$ which comes from Fig. 2 multiplied by an estimated section area from Whitehead and Worthington. $Q_{1\text{current meters}}$ really does not enter except to determine that area. The geostrophic results use an effective volume flux which is given by $Q_{2\text{geostrophic}} = Q_{1\text{geostrophic}} \times [Q_{2\text{current meters}}/Q_{1\text{current meters}}]$. This flux is of the order of the steady geostrophic flux and is thus consistent with everything known to date about the flow over the sill. The numbers in columns 5 and 6 in Tables 1a, b can be inserted into Eq. (6) and for the current meter estimates the following formulas result:

$$h'/H = 0.0057 \sin(\omega t - 90) \quad (7)$$
$$= 0.019 \sin(\omega t - 89) \quad (8)$$
$$= 0.058 \sin(\omega t - 87) \quad (9)$$

for water colder than 1.3°, 1.2° and 1.1°C, respectively. For the geostrophic estimates the following formulas are found:

$$h'/H = 0.028 \sin(\omega t - 89) \quad (10)$$
$$= 0.19 \sin(\omega t - 85) \quad (11)$$
$$= 0.99 \sin(\omega t - 78). \quad (12)$$

The coefficients in Eqs. (7)–(10) would lead to predictions of h' much smaller than 50 m (given H of order 200–500 m), and scatter with less than 50 m would be undetected from background ocean turbulence. However, those in (11) and (12) are large enough

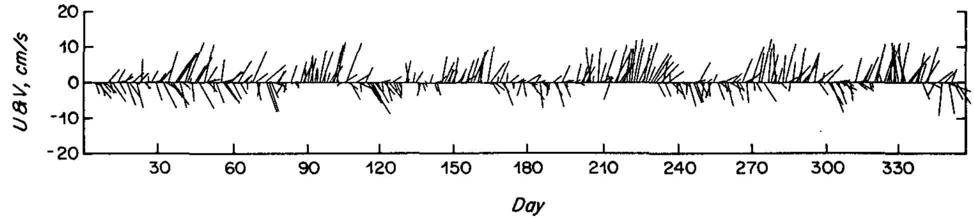

FIG. 2. The current meter record from the bottom current meter in the west mooring at 4°N. This was in water colder than 1.2°C and is shown in Fig. 8d of Whitehead and Worthington (1982).

to warrant further calculation. The final step is to arrive at a value of h' by estimating H. For the water colder than 1.1° we use estimates of volume and area in Table 3 of Whitehead and Worthington (1982). Here H is found by taking twice the volume of the 1.0° to 1.1°C layer (2 times 2.2×10^3 km^3) divided by the area of the 1.1° isotherm (55×10^{13} cm^2). The factor of two comes from approximating the bathymetry as a V-shaped trough, so that the volume is given by $AH/2$. This gives $H = 80$ m for the 1.1°C isotherm. In a similar way we find $H = 117$ m for the 1.2°C isotherm.

This yields predictions for the total estimated excursions of the isotherms $2h'$ which are shown in column 7 of Table 1. One prediction, whose verification can be sought with ocean data, is that the geostrophic flux model predicts a total vertical excursion of *158 m* during the 60-day period for the 1.1° isotherm. A second prediction which could be tested by ocean data is a total vertical excursion of *44 m* over the 60-day period for the 1.2°C isotherm. Other predictions of h' are too small to be detected in the ocean.

Other basin geometries give even bigger predictions. For $n = 1$, we get, instead of (11),

$$\left(\frac{h'}{H}\right)^2 = 0.38 \sin(\omega t - 85), \qquad (13)$$

and instead of (12),

$$\left(\frac{h'}{H}\right)^2 = 2.10 \sin(\omega t - 78). \qquad (14)$$

This gives total estimated excursions of *231 m* for the 1.1°C isotherm and *62 m* for the 1.2°C isotherm. Larger values of n give approximately these numbers, too.

It was felt that vertical excursions of isotherms with an amplitude over roughly 50 m might be seen as scatter in the historical data set. The next section describes a review of the historical hydrocast data.

4. Observations of variation of the 1.1° and 1.2° isotherm

Is there any evidence for an excursion as large as 160–230 m of the 1.1°C isotherm or 45–60 m of the 1.2°C isotherm? To determine whether the 1.1° or 1.2°C isotherm can be mapped to a precision which would reveal a 60-day excursion with the amplitudes predicted above, data have been gathered from all known hydrocasts north of 4°N in water colder than 1.2°C. The included data are from the western Atlantic GEOSECS, the tropical Atlantic leg of the Transient Tracers in the Ocean (TTO) experiment, *Oceanus* cruises 36 and 52, *Crawford* cruise 165 and *Conrad* cruise 29-06. Locations are shown in Fig. 3. Stations with water colder than 1.2°C extend 1400 km northwest of the 4°N sill.

Data on the depth of the 1.1 and 1.2°C isotherm for the stations shown on Fig. 3 are given in Table 2, which also reports position and time information and linear distance (measured from Fig. 3) from the sill at 4°N. Isotherm depths (Fig. 4) show that the 1.1°C isotherm does not exhibit a vertical scatter of 160–230 m which our model has predicted using the geostrophic estimates. There is also no scatter in the depth of the 1.2°C isotherm of 45–60 m. Moreover, the apparent scatter

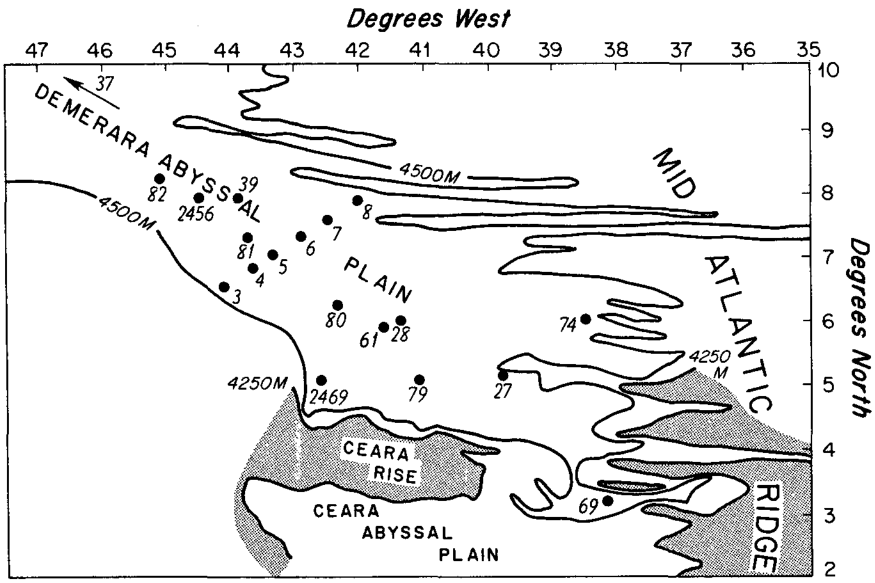

FIG. 3. Locations of stations listed in Table 2. These reveal water colder than 1.2°C in the western Atlantic north of 4°N. Also shown is a station location from TOGA 69 (Katz, personal communication) at 3°10′N, 38°28′W.

TABLE 2. Depth of the 1.2° and 1.1°C isotherms north of 4°N. Positions and date of stations are also given. Stations 3–8 are from *Oceanus* cruise 36. Stations 79–82 are from *Oceanus* cruise 52. Stations 37 and 39 are from GEOSECS. Stations 2456 and 2469 are from *Crawford* 165. Stations 61 is from the Transient Tracer in the Ocean's Tropical Atlantic Study. Station 74 is from a TOGA cruise (Eli Katz, personal communication).

Station	Distance (km)	Depth 1.2°C	Depth 1.1°C	Depth Bottom	Longitude (deg)	Longitude (min)	Latitude (deg)	Latitude (min)	Date (d, mo, yr)
27	120	4400	4540	4680	5	9	39	48	10/12/77
79	200	4430	4540	4680	5	7	41	6	6/12/78
74	210	4510	None	4644	5	59	39	24	10/06/88
28	270	4421	4610	4700	6	3	41	11	10/12/77
61	275	4560	4660	4702	5	58	41	21	3/01/83
2469	325	4460	4520	4715	5	12	42	24	24/02/68
80	360	4470	4570	4690	6	14	42	22	7/12/78
7	465	4569	none	4819	7	31	42	33	30/11/77
8	465	4780	none	4836	7	48	42	00	1/12/77
6	470	4592	4780	4788	7	19	42	53	30/11/77
5	480	4534	4695	4786	7	4	43	21	30/11/77
4	500	4516	4646	4753	6.	45	43	37	29/11/77
3	520	4539	none	4642	6	32	44	5	29/11/77
81	535	4540	none	4700	7	10	43	46	7/12/78
39	580	4610	none	4793	7	58	43	51	17/10/72
2456	610	4632	none	4644	7	49	44	21	20/02/68
82	700	4550	none	4625	8	10	45	6	8/12/78
37	1400	5006	none	5073	12	18	50	59	13/12/72

in this figure appears to be systemically connected to a cross-channel (geostrophic?) variation. For instance, stations 8 and 61 are furthest to the east and they clearly have the deepest 1.1°C isotherms. The cross-channel tilt is shown in Fig. 5 which shows a cross section of the *Oceanus* 36 data which was taken within a three-day interval. The scatter which appeared in Fig. 4 for depths of both isotherms in stations 3–8 is clearly due to a cross-channel tilt. Also shown are the 1.2°C isotherm depths from *Oceanus* 52 casts 81 and 82 which were taken one year later, and from GEOSECS cast 39 which was taken five years earlier, and they are at

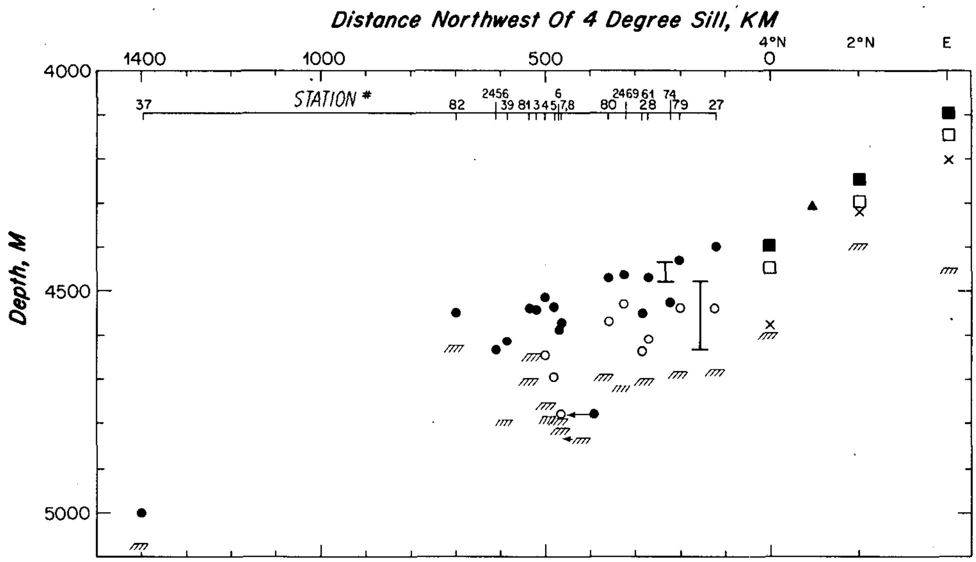

FIG. 4. A section along the deepest passageway into the North Atlantic from the Equator to approximately 12°N. This shows all known observations of the depths of the 1.2°C (closed circles), 1.1°C (open circles), and 1.0°C (X) isotherms north of 4°N and some data from 0 to 4°N as well. The average depths of the 1.2°C isotherm (closed squares), the 1.1°C (open squares) isotherm, and the 1.0°C isotherm (x) are from Figs. 4–6 of Whitehead and Worthington. The solid triangle is the depth of the 1.2°C isotherm (Katz, personal communication) from TOGA 69. Error bars showing 158 and 44 m-depth excursions as predicted by Eqs. (12) and (11) are also shown.

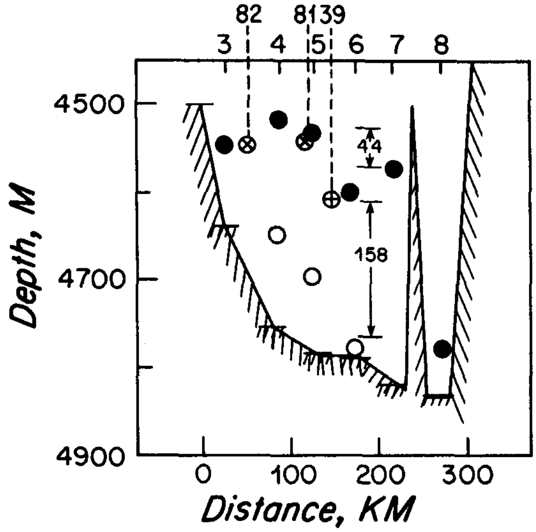

FIG. 5. Depth of isotherms for all stations between 460 and 610 km from the sill. Lateral position is plotted as a function of the distance across channel from the 4500 m isobath in Fig. 3. For stations 3–8 the 1.2°C isotherms are solid circles and the 1.1°C isotherms are open circles. The 1.2°C isotherms from *Oceanus* 52 stations 81 and 82, which were taken one year later are shown as ⊗s. The 1.2°C isotherm from GEOSECS cast 39, which was taken five years previously, is shown as a circled plus. The vertical distances of 158 and 44 m, which were predicted excursions of the isotherms by equations (12) and (11) are shown.

approximately the same depth. Thus temporal variation of the depth of the 1.1°C isotherm of the order of 160–230 m is clearly not indicated by the present data. Temporal variations of the order of 45–60 m by the 1.2°C isotherm cannot strictly be ruled out but also appear to be unverified.

5. New volume estimates

Estimates for volume of the water masses in Whitehead and Worthington were taken from work sheets prepared from data available in 1980. The hydrocast data used were the *Oceanus* data shown on Table 2, plus the IGY data. Bathymetry was obtained from existing charts. This sparse data may lead to considerable errors and we now estimate that those estimates for area and volume (which we used in the calculations in Section 3) may be incorrect by up to a factor of almost 4. To illustrate this, Fig. 6 shows two new maps of the depth of 1.1° and 1.2°C contours north of 2°N in the western North Atlantic. These are drawn from the data in stations shown by dots. The stations north of 4°N are given in Table 2. The stations at 4°N and further south are from *Oceanus* cruises 36 and 52, and the depths of the isotherms are taken directly from Figures 4 and 6 of Whitehead and Worthington. The position where the isopotential temperature surfaces intersect the ocean bathymetry were determined using the more recent bathymetric chart of Moody, Hayes, and Connary (1979). The tongues of 1.1° and 1.2°C waters from the Worthington and Wright atlas are shown as dashed curves and bear little detailed relation to the new outline. The area of the 1.1° isotherm is 197×10^{17} cm^2 (in contrast to 55×10^{17} in Whitehead and Worthington) and the area of the 1.2°C isotherm is 514×10^{17} cm^2 (in contrast to 191×10^{17}). It is clear that advances in the bathymetry and in the hydrographic data base since 1960 produce a map with greatly improved resolution. Volumes based upon these new data are significantly different from previous ones.

Even these maps are incomplete and may lead to large absolute errors if taken at face value. For example, in Figure 6a there are no stations over a vast area in the east. The bathymetry there is not contoured in present maps in enough detail to determine either the true thickness of the water column or its true areal extent. In Fig. 6b, the entire northern extent of the tongue is based on GEOSECS station 37 which displays results that are very different from results shown as stars in Figure 6b from stations 20 and 22 of leg 1 of the Transient Tracers of the Ocean's (TTO) Tropical Atlantic Study (Williams 1986). Those two casts did not even record water as cold as 1.3°C. The station denoted by the northernmost star recorded $\theta = 1.31$°C at 5066 m, and the other one recorded $\theta = 1.414$°C at 4862 m. Whether there is a temporal variation which only the GEOSECS cast caught, or whether one or the other of the stations has a systematic calibration error, is unclear.

6. Consequences of the new volumes

Using the newly computed areas, Table 3 was constructed. This is an improved version of Table 1. Only the geostrophic results will be presented, since only they are large enough to contradict the historical data. The new parameters in columns 2, 4 and 5 are determined by multiplying by the ratio of new area estimates to historical ones. The estimate of $2h'$ in column 7 is found by dividing by this number. It shows that the predicted excursions of the interfaces are under 50 m. Thus geostrophic results cannot be ruled out on the basis of this model.

Of more fundamental importance is the fact that these new volumes lead to new estimates of the steady residence times, and therefore heat and salt diffusivities. For this, we assume that Whitehead and Worthington's estimate for the thickness of the waters is correct. Old and new values for these properties are given in Table 4. We now note that some of the revised estimates of diffusivity from the current meter estimates are extremely small—possibly to the extent that they could be ruled out for other reasons. The lowest value of 0.027 cm^2 s^{-1} is close to, but still greater than, the molecular values of water of .00134 cm^2 s^{-1}. We are not aware of any convincing argument that would support discarding them. Even the largest values are now

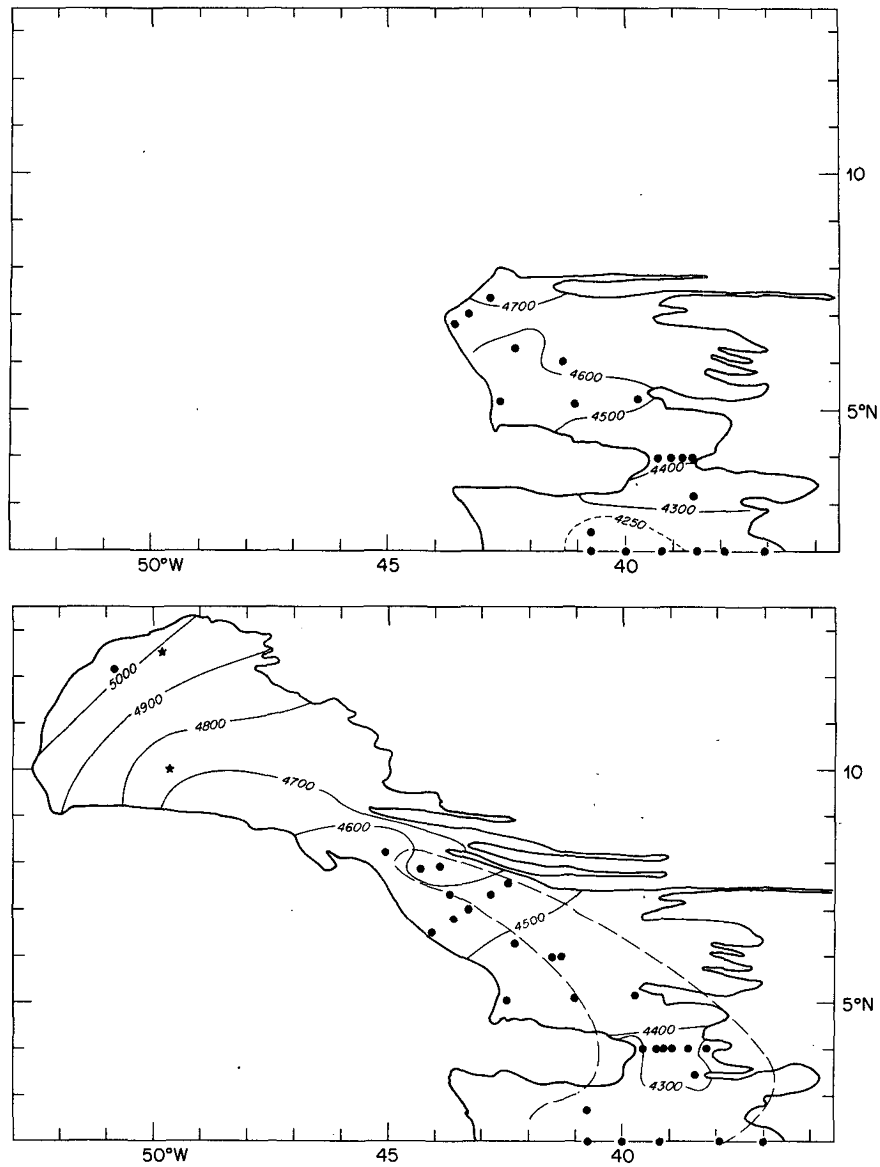

FIG. 6. Maps of the tongues of (a) the 1.1°C isotherm and (b) the 1.2°C isotherm north of 2°. The contours show estimated depths and the dashed lines show the corresponding tongue in Worthington and Wright. The two stars denote stations 020 and 022 of the Transient Tracers in the Ocean's Tropical Atlantic Study. They revealed bottom water warmer than 1.3° at depths considerably greater than those shown.

considerably less than the values of 3 $cm^2\ s^{-1}$ found by Hogg et al. (1982) and Saunders (1987). They all used similar techniques to deduce mixing coefficients in other basins. They may have inaccuracies in their estimate of the volumes of their deepest waters similar to the ones we had.

TABLE 3. Some parameters and the predicted vertical excursion of isotherms $2h'$ for water colder than 1.2°C in the North Atlantic using new volume estimates from the new areas revealed by Figs. 6a and 6b.

1 Temperature (°C)	2 Volume (10^{12} m^3)	3 Volume flux upward (10^{12} m^3 d^{-1})	4 T (days)	5 ωT	6 Q_2/Q_1	7 $2h'$ (m)
1.2	30.14	0.096	315	32	2.2	16.4
1.1	7.88	0.050	158	16.5	4.5	44

TABLE 4. Revised estimates of properties of water colder than 1.2°C based on the new areas revealed by Figures 6a and b. These are compared with earlier estimates by Whitehead and Worthington (W&W).

Isotherm (°C)	Area (10^{13} cm^2)		Factor change	Residence time Y		Heat mixing coefficients (cm^2 s^{-1})		Salt mixing coefficients (cm^2 s^{-1})	
	W&W	New		W&W	New	W&W	New	W&W	New
				Current meters					
1.2	191	514	2.69	2.5	6.7	0.186	0.069	0.34	0.13
1.1	55	197	3.58	1.9	6.8	0.095	0.027	2.12	0.06
				Geostrophic estimates					
1.2	191	514	2.69	0.3 (94 days)	0.7 (254 days)	4.010	1.12	4.183	1.55
1.1	55	197	3.58	0.1 (44 days)	0.4 (157 days)	3.327	0.93	3.349	0.94

7. Concluding remarks

A time-dependent model is developed for flow of Antarctic Bottom Water over the Ceara Abyssal Plain at approximately 4°N into the North Atlantic. Using values from Whitehead and Worthington for mixing coefficients and flow fluctuations, the model predicts vertical excursions of the 1.1°C isotherm of 160–230 m, and of 45–60 m for the 1.2°C isotherm. Historical data did not show evidence of such large excursions. A new study reveals that the estimates in Whitehead and Worthington of areas covered by the 1.1° and 1.2°C isotherms are too small by factors of 2 to 4 due to the sparse hydrographic and bathymetric data available then. This leads to smaller predicted vertical excursions of the isotherms. These are consistent with observations. New diffusivities are calculated based on the new areas and are given in Table 4. They are smaller than the previous estimates by the above factors. Although it was hoped that this study would resolve the disagreement between the current meter and geostrophic estimates for volume flux of water colder than 1.2°C into the North Atlantic, the disagreement remains.

Acknowledgments. Research on nonlinear jets has been funded by the Office of Naval Research under contract N00014-87-K-0007, NR 083-004 Division 112/CS. I want to thank Larry Pratt for interesting discussions on this problem, and Eli Katz for supplying me with TOGA data in the area of interest.

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