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Tracing Lower North Atlantic Deep Water Across the Equator

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Water whose ultimate source is the overflow at the Denmark Strait can be traced using maps of salinity, dissolved oxygen, and silica concentration on deep density surfaces south along the American continent to at least 5°S. Several property extrema whose resolution is at the limit of accuracy of the measurements occur on these density surfaces, which nevertheless may indicate areas of stronger mixing along the crest of the Mid-Atlantic Ridge. Transport estimates for that part of the southward flow within 600 km of the western boundary near 10°N suggest that $10 + 3 \times 10^6$ m^3 s^{-1} of Lower North Atlantic Deep Water moves south there. The total southward deep water transport near 10°N is estimated to be close to 25×10^6 m^3 s^{-1}.

1. INTRODUCTION

Dense water formed in the Nordic seas between Iceland, Scotland, and Greenland spills out across the Denmark Strait and the Iceland-Scotland sills to form bottom water at the northern end of the Atlantic Ocean. *Swift* [1984] recently discussed property distributions near the sills and has traced this overflow water to 30°N. *Jenkins and Rhines* [1980] presented evidence for the southward flow of this water in a narrow (50 km wide) western boundary current at 30°N on the Blake-Bahama Outer Ridge. While this water has been shown to flow as far south as 13°N [*Fine and Molinari*, 1988], the expected continuation of this current into the South Atlantic has not been evident in previous work, despite the unambiguous presence of this water south of the equator [e.g., *Warren*, 1981]. Clear indications that some of this water must penetrate farther south are also apparent in the properties of water near 4000 m along the equator [*Bainbridge*, 1980], which show the characteristics of overflow water: relatively high oxygen concentrations and low nutrient concentrations.

A notable characteristic of the tritium core used in the *Jenkins and Rhines* [1980] study to identify the overflow water is its potential temperature, 1.9°C. *Olson et al.* [1986] also found tritium at this temperature farther south along the boundary at 26°N and 22°N, with maximum values occurring at a potential temperature of 2.5°C. It may be anticipated, then, that the characteristics of water on deep isopycnal surfaces near these temperatures vary between those of northern and southern origin, and their distribution should provide an indication of the southward penetration of the overflow water. Several properties on deep density surfaces are presented here to demonstrate the continuation of this boundary current into the South Atlantic. The density surfaces chosen are roughly 3500 m and 4200 m deep and span the high freon concentrations illustrated by *Fine and Molinari* [1988] at 13°N. The deeper of the two surfaces usually lies between 1.8°C and 1.9°C in the western trough of the North Atlantic, while the shallower surface is closer to 2.2°–2.26°C and intersects a part of the tritium maximum at 22°–26°N [*Olson et al.*, 1986].

Wright [1970], using geostrophic transport calculations with International Geophysical Year (IGY) data determined that about 9×10^6 m^3 s^{-1} of North Atlantic Deep Water flows south at 16°N, 8°S, and 32°S. He did not distinguish lower from upper components, although these have different North Atlantic origins and thus merit individual attention. Transport calculations using the more recent data are being prepared separately, but some preliminary results are presented here to indicate the size of the flow. This is of importance not only for the long-range influence of deep water, but also for its local effects: south of the region where the overflow water lies on the bottom, it receives directly the upwelling Antarctic bottom water, buffering, in a sense, the middepths from the cooler, fresher water of southern origin.

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2. DATA

Most of the temperature and salinity data used here (Table 1) were processed by the Woods Hole Oceanographic Institution (WHOI) conductivity-temperature-depth (CTD) group using their standard practices [*Millard*, 1982]. The expected accuracy of the measurements is ±0.002°C, ±0.002‰, and ±2 dbar for temperature, salinity, and pressure.

Oxygen and nutrient measurements were also made on the WHOI cruises. These data are expected to be accurate to ±0.05 mL/L for oxygen and 1% for the nutrients (only silica is used here).

Other data used here are from the Geochemical Ocean Section Study (GEOSECS) program (July 1972 to April 1973 for Atlantic stations [*Bainbridge*, 1980]) and the Transient Tracers in the Ocean (TTO) program (North Atlantic study, April 1981 to October 1981; Tropical Atlantic study, December 1982 to February 1983 [*Brewer et al.*, 1985]). The accuracy of these data is not reported but is expected to be similar to the WHOI measurements. A few values from the *Levitus* [1982] atlas are used where noted.

Salinity values were corrected for differences between standard seawater batches used for calibration according to *Mantyla*'s [1987] Table 2. The corrections (ΔS) are displayed

TABLE 1. Data Sources and Correction to a Uniform Standard (P93).

Location	Time	Ship and Cruise	Batch	ΔS
36°N	June 1981	*Atlantis* II 109 leg 1	P81	+0.001
24°N	Aug. 1981	*Atlantis* II 109 leg 3	P81	+0.001
13°N	Jan. 1983	*Oceanus* 133 legs 2	P90	−0.002
23°S	Feb. 1983	*Oceanus* 133 leg 4		
11°S	March 1983	*Oceanus* 133 leg 5	P93	0
53°W	April 1983	*Oceanus* 133 leg 7	P93	0
35°W	July 1983	*Knorr* 104	P93	0
64°W	April 1985	*Endeavor* 129	P93	0
TTO	1981–1983	*Knorr*	P80	+0.004
GEOSECS	1972–1973	*Knorr*		

with batch numbers in the table given here. During the TTO program, both standard seawater batch P80 and batch P90 were used to calibrate the salinometer measurements, but the reported data are referenced to batch P80 [*Mantyla*, 1987]. A mean offset on a deep isopycnal surface between WHOI data and the TTO data of ±0.004‰ (with a standard deviation of 0.002‰) was found in places where they overlapped, so this value was added to TTO data used here in agreement with *Mantyla*'s [1987] differences. (Note that any discrepancy between these data sets is slightly exagerated on a deep isopycnal surface because for nearly linear Atlantic deep *T–S* curves the salinity offset is somewhat greater along a curve of constant density than it is along the salinity axis. To illustrate the magnitude of this difference, the standard deviation of the 24°N eastern basin CTD data was computed relative to *Saunders*' [1986] reference θ–S relation. The result was a standard deviation of 0.002‰ along the S axis and 0.003‰ along constant σ_4 curves.) These corrections improved the consistency of salinity measurements between the different data sets. A comparison of GEOSECS salinity with TTO salinity in nearby stations on an isopycnal surface revealed a standard deviation of 0.008‰ of the difference. Some of this variability may be real, from changes in time and spatial gradients; however, the better agreement between the other data sets suggests some other problem, and the GEOSECS salinity was not used here.

Not all stations in a given cruise were used for making the horizontal maps of properties on constant density surfaces because of computer time and cost constraints. This was justified by a few plots that were made with all the stations, which did not show any new structure. The locations of missing stations are easily visible in the maps as portions of cruise track with unusual spacing.

3. DISTRIBUTION OF SALINITY, OXYGEN, AND SILICA

Plots of salinity of isopycnal surfaces are used to examine variations in the θ–S relation, since advection occurs primarily along a constant density surface. Most of the variation of salinity is expected to be caused by the mixing of salty deep water from the northern North Atlantic with fresh bottom water from the South Atlantic. Other possible causes of variation include vertical mixing along the boundaries such as the Mid-Atlantic Ridge, perhaps caused by local heating or mechanical mixing. The speed required of deep flow to make mechanical mixing important, several tens of centimeters per second, seems unrealistically high except perhaps along the western boundary.

Deep isopycnal surfaces σ_4 = 45.85 kg m^{-3} (Figure 1) and σ_4 = 45.90 kg m^{-3} (Figure 2; calculated using the 1980 equation of state) have been chosen to emphasize the Lower North Atlantic Deep Water (LNADW). Also displayed on both surfaces are oxygen concentration and dissolved silica. These densities correspond roughly to a potential temperature of 2.2°C and 1.9°C, and to depths of 3500 m and 4200 m in the western trough. The total variation of salinity on these surfaces is small, about 0.04‰, and it was necessary to make corrections owing to salinity differences in the standard seawater used to calibrate the salinometers, which in turn are used to calibrate the CTD profiles [*Mantyla*, 1987]. The corrected data are presented.

In the western trough the basic pattern of the salinity distribution on both surfaces is of relatively salty water in the north and west and fresh water in the south and east. High-salinity water extends south along the western boundary past the equator, and throughout this path of over 55° of latitude the salinity varies by only about 0.005‰ (Figure 1*a*). South of 5°S the salinity decreases sharply by more than 0.01‰ to values less that 34.91‰ at 11°S. This salinity distribution confirms, at slightly better resolution, the picture provided by *Worthington and Wright* [1970] at 2.2°C and 1.9°C. In addition to being relatively salty, the water in the west is more highly oxygenated (Figures 1*b* and 2*b*) and has a lower dissolved silica concentration (Figures 1*c* and 2*c*), which is also characteristic of Lower North Atlantic Deep Water. These property distributions all result from the southward flowing deep western boundary current.

In addition to the tracer evidence, the continuity of this current is becoming apparent from recent direct measurements as well. For example, yearlong direct current measurements near 8°N at the western boundary by *Johns et al.* [1990] show the presence of a rather uniform southwestward flow above the continental rise at 4600 m depth. The mean speed of the near-bottom flow over the measurement period was about 20 cm s^{-1}.

Along the western boundary near 10°N a salinity maximum appears on the lower, 45.90 kg m^{-3} isopycnal. This isolated stretch of high salinity is correlated with low silica values as well (Figure 2*c*). Although oxygen concentration shows no particular high there (Figure 2*b*), individual station values do continue to increase toward the western boundary up to 6.05 mL L^{-1} near 10°N, and somewhat higher near 7°N. In general, vertical mixing between bottom water and deep water should on the contrary reduce salinity and oxygen concentration in the LNADW. This isolated feature therefore seems to be an effect of sampling: inadequate spatial resolution or time-dependent property distributions.

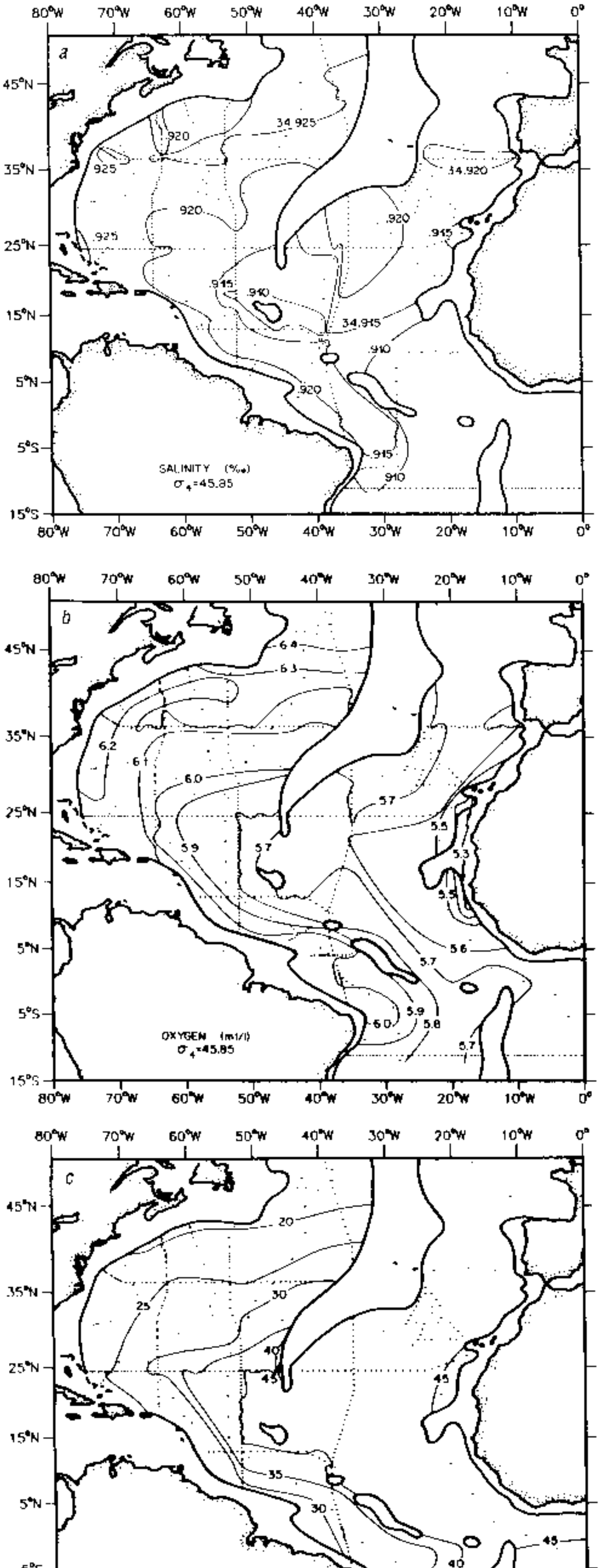

Fig. 1. (*a*) Salinity (per mil), (*b*) oxygen (milliliters per liter), and (*c*) silica (micromoles per liter) on $\sigma_4 = 45.85$ kg m^{-3} (potential density anomaly referenced to 4000 dbar). Topography adapted from *Worthington and Wright* [1970].

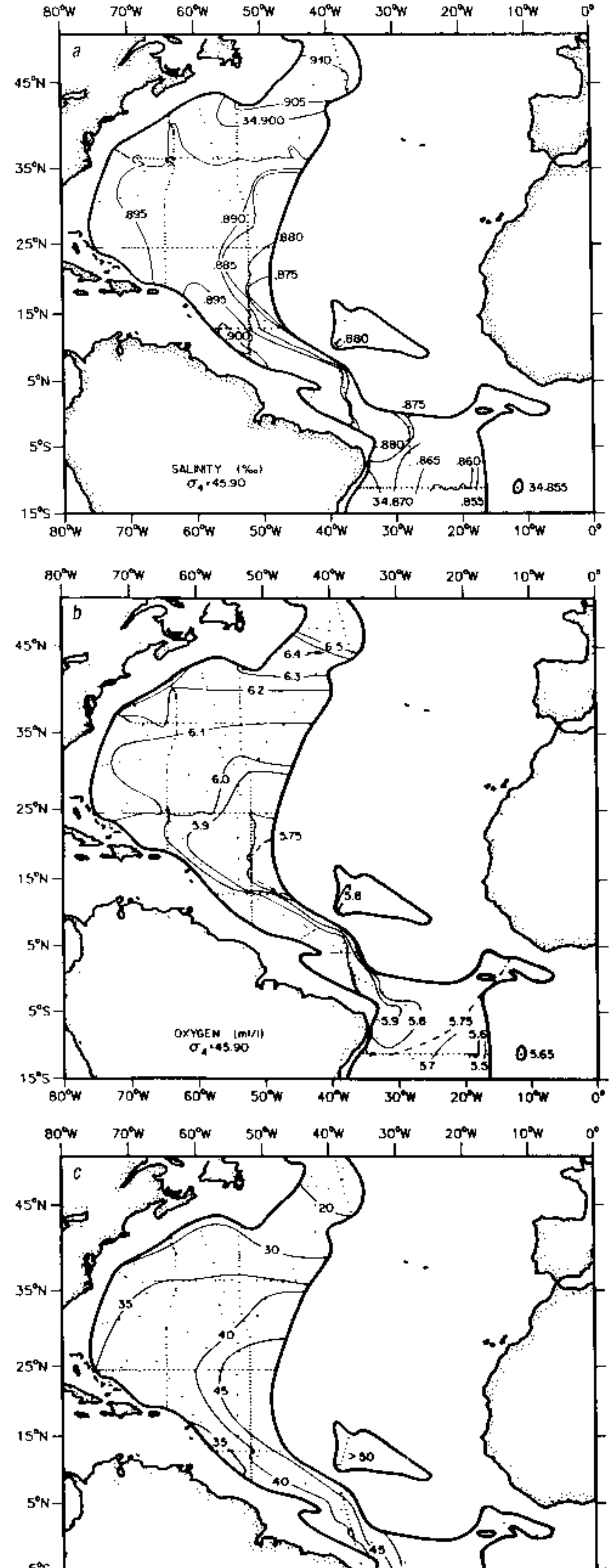

Fig. 2. (*a*) Salinity (per mil), (*b*) oxygen (milliliters per liter), and (*c*) silica (micromoles per liter) on σ_4 = 45.90. Values in the southeastern part of the eastern trough in Figure 2*b* are from the Levitus atlas.

TABLE 2. Volume Transport Estimates Close to the Western Boundary

Latitude	AAIW 400–1200 m	UNADW 1200–2900 m	Stations	LNADW 2900–4400 m	Stations (Width, km)	DW Sum	IW– DW Sum
13°N	4.3	−16.8	2–6	−8.4	4–11 (650)	−25.2	−20.9
7°–10°N	13.1	−12.6	215–221	−13.2	218–226 (450)	−25.9	12.8

Units are 10^6 m^3 s^{-1}, positive is north. AAIW is Antarctic Intermediate Water, UNADW is Upper North Atlantic Deep Water, and LNADW is Lower North Atlantic Deep Water. DW is the sum of the deep water transports, IW + DW includes intermediate water.

Near the Mid-Atlantic Ridge, water is less saline. Fairly uniform property values above the eastern flank of the ridge (Figure 1) are consistent with a northwest flow there, suggested by *Mann et al.* [1973] on the basis of both silica profiles and also the salinity distribution on potential temperature surfaces produced by *Worthington and Wright* [1970]. *Mann et al.* [1973] remarked on the uniformity of properties across the ridge from east to west between 2.4° and 2.6°C, but this clearly applies to somewhat deeper levels as well. This raises the issue of a deep western boundary current in the eastern trough, which according to *Stommel and Arons'* [1960] prescription ought to flow north up to mid-latitudes where it reverses as part of the recirculation at the poleward end of a closed basin. The Mid-Atlantic Ridge is a very uneven boundary, though, and it is unclear whether or not the model should be applied, not to mention the ill-defined location and strength of a southern source.

The only other way salinity could be decreased near the ridge is through vertical mixing, producing a downward flux or sink there. On the upper density surface in the latitude range of roughly 10°N to 20°N, there is a salinity minimum that is isolated from surrounding water by about a 0.005‰ change in salinity. A minimum of similar amplitude also appears on the lower surface (Figure 2*a*). In each case the extremum is defined by several stations from two different cruises. Furthermore, where other ship tracks cross the 13°N section, the salinities agree. Thus shifting one section or another by 0.005‰ to remove the extremum would not seem to be justified.

On the deeper surface the distribution of oxygen has the same basic shape as the salinity, with low oxygen correlated with low salinity (Figure 2*b*). The oxygen distribution supports the isolation, but again the amplitude (from western trough to eastern trough) is marginal, about 0.05 mL L^{-1}. Finally, on the upper surface, both oxygen and silica are extreme at the eastern boundary. On this surface, eastern boundary sources and sinks together with lateral advection and diffusion seem to overwhelm any potential vertical mixing effect.

These Mid-Atlantic Ridge extrema, if real, would be direct evidence of mixing between the Antarctic Bottom Water and Lower North Atlantic Deep Water and may represent the regions of strongest vertical mixing along the crest of the Mid-Atlantic Ridge. However, they can only be regarded at this point as a hint of such an effect, whose large-scale signal is weak compared to that of the basic flow.

4. GEOSTROPHIC VOLUME TRANSPORT

Geostrophic transport calculations were made using some of the stations near the western boundary to estimate the strength of the southward flow there. The general northward flow of Antarctic Bottom Water and Intermediate Water below and above North Atlantic Deep Water suggests two possible reference levels, one at about 4000 m depth and the other at roughly 1000 m depth. To estimate the northward flow of bottom water, *Wright* [1970] chose a zero velocity surface near 2°C, or about 4000 m depth in the interior, but shallower near the boundary by several hundred meters. While this choice is acceptable for bottom water transport estimates, it is clearly wrong for those of deep water, since water of this temperature is flowing south along the boundary north of the equator.

A shallower reference level choice, on the other hand, produces the correct sense of intermediate and deep flow at the boundary. The most important guide to the upper reference level choice is the salinity distribution, which has a minimum in the Intermediate Water, above a maximum in deep water. A reference level of 1200 m depth (or the bottom, where shallower) seemed to be best as a uniformly valid choice throughout the latitude range of the sections which lies between the two extrema, and this value was used in all calculations (Table 2). This choice is meant to be valid only in the restricted latitude range shown below and only close to the boundary. Bottom triangle transports were approximated by multiplying the deepest velocity common to a station pair by the area below the deepest common level.

To allow comparison, the deep water was divided into both upper and lower components. The contribution of Antarctic Intermediate Water was also estimated, but Antarctic Bottom Water transport is not considered here. Farther offshore, new reference level choices are required to arrive at basin-wide estimates. The depth range defining the components and calculated transports are given, together with their sum showing the net southward movement of water in each depth range (Table 2). The error implied by moving the reference level down by several hundred meters amounts to less than 10%. Moving it up into Intermediate Water changes the estimates substantially. In addition, these estimates suffer from a inexact definition of the offshore edge of the current. The 6.0 mL L^{-1} contour is used here as the edge (Figure 2*b*); variations from one station pair to another at the offshore edge can amount to 50%.

Considering the result at 7°–10°N and 13°N, the calculation shows a southward LNADW transport of about $10 \pm 3 \times 10^6$ m^3 s^{-1}. The deep water sum is nearly the same, 25×10^6 m^3 s^{-1}, at both sections and is almost triple the estimate of *Wright* [1970] and *Fine and Molinari* [1988], probably because of their deep reference level (and therefore weaker deep velocities). The wider station spacing of the data used by *Wright* [1970] must play a role too, but *Roemmich*'s [1983] inverse method transport calculation using the same data with an initial reference level choice at 2000 m resulted

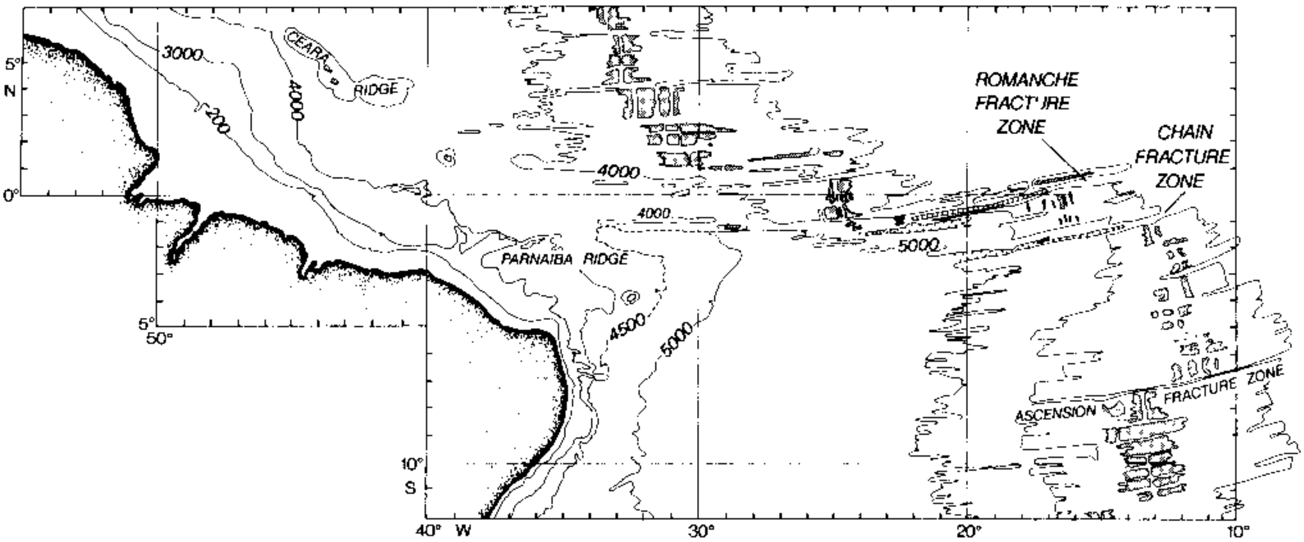

Fig. 3. Bathymetry of the western equatorial Atlantic region, simplified from GEBCO chart 5-12 of the South Atlantic Ocean. Mid-Atlantic Ridge areas shallower than 3000 m depth are shaded, tick marks indicate deeps.

in a LNADW (similarly defined) transport of 15.1×10^6 m^3 s^{-1} to the south at 8°N. This result, though, seemed to depend more on the mass conservation constraint in deep layers than on the 8°N hydrography. His combined deep water transport was 23.7×10^6 m^3 s^{-1} to the south at 8°N, and by construction the same figure obtains at 8°S, 24°S, and 24°N too.

The division of the deep water into depth intervals ignores the tendency for water to flow along isopycnals. Alternatively, transports could be divided into different temperature or density classes to define the various components, but the substantial change of temperature, salinity, and other tracers with latitude on the deep density surfaces show such a division to be arbitrary and problematic as well. Evidently, some kind of lateral discrimination could be made when choosing boundaries between water masses, in addition to the usual vertical divisions.

5. CONCLUSION

The greatest change in the salinity of Lower North Atlantic Deep Water along the western boundary on two deep isopycnals occurs between 5°S and 11°S. This indicates that the Lower North Atlantic Deep Water current has crossed the equator without substantial modification. The modification of its tracer characteristics south of the equator, in the Brazil Basin, is understandable because here the Antarctic Bottom Water is a western boundary current too, and so the two currents lie next to one another.

Geostrophic transport calculations obtained using a reference level choice of 1200 m next to the western boundary support the tracer evidence of the southward extension of the deep western boundary current across the equator.

The oxygen distribution in particular supports the picture of cross-equatorial flow which branches in the Brazil Basin, some continuing south and some turning east. The oxygen distribution south of the equator is aligned with an eastward extension of the continental rise near 3°S, east of the Parnaiba Ridge (Figure 3). This topographic feature seems too broad and deep to be much of a guide toward the entrance to the Chain Fracture Zone at 2°20′S or the entrance to the Romanche Fracture Zone at 0°45′S. Nevertheless, the LNADW water does make its way across the Brazil Basin and Mid-Atlantic Ridge, as the high oxygen concentrations at 4000 m depth (and 45.87 kg m^{-3}) on the equator east of the Chain Fracture Zone and Romanche Fracture Zone exits attest [*Bainbridge*, 1980].

Although the high oxygen concentrations east of the Mid-Atlantic Ridge occur near the equator, that water certainly cannot have moved east precisely along the equator because this path is blocked by the ridge itself near 25°W (Figure 3). Moreover, the two main fracture zones are tilted with respect to the zonal direction, and water therefore enters them at the off-equatorial latitudes mentioned above. All these factors support a path for the lower deep water somewhat south of the equator and suggest that the more southerly Chain Fracture Zone may be more of a guide than was previously thought.

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REFERENCES

Bainbridge, A. E., *GEOSECS Atlantic Expedition*, vol. 2, *Sections and Profiles*, 196 + xiv pp., National Science Foundation, Washington D. C., 1980.

Brewer P. G., J. L. Sarmiento, and W. M. Smethie, The Transient Tracers in the Ocean (TTO) program: The North Atlantic study, 1981; the tropical Atlantic study, 1983, *J. Geophys. Res.*, *90*, 6903–6905, 1985.

Fine, R. A., and R. L. Molinari, A continuous deep western boundary current between Abaco (26.5°N) and Barbados (13°N), *Deep Sea Res.*, *35*, 1441–1450, 1988.

Fuglister, F. C., *Atlantic Ocean Atlas of Temperature and Salinity Profiles and Data From the International Geophysical Year of 1957–1958*, *Woods Hole Oceanogr. Inst. Atlas Ser.*, vol. 1, 209 pp., Woods Hole Oceanographic Institution, Woods Hole, Mass., 1960.

Jenkins, W. J., and P. B. Rhines, Tritium in the deep North Atlantic Ocean, *Nature*, *286*, 877–880, 1980.

Johns, W. E., T. N. Lee, F. A. Schott, R. J. Zantopp, and R. Evans, The North Brazil Current retroflection: Seasonal structures and eddy variability, *J. Geophys. Res.*, *95*(C12), 22103–22120, 1990.
Levitus, S., Climatological atlas of the world ocean, *NOAA Prof. Pap. 13*, 173 pp., U.S. Govt. Print. Off., Washington, D. C., 1982.
Mann, C. R., A. R. Coote, and D. M. Garner, The meridional distribution of silicate in the western Atlantic Ocean, *Deep Sea Res.*, *20*, 791–801, 1973.
Mantyla, A. W., Standard seawater comparisons updated, *J. Phys. Oceanogr.*, *17*, 543–548, 1987.
Millard, R. C., Jr., CTD calibration and data processing techniques at WHOI using the 1978 practical salinity scale, paper presented at International STD Conference and Workshop, San Diego Sect., Mar. Technol. Soc., La Jolla, Calif., Feb. 1982.
Olson, D. B., G. H. Ostlund, and J. Sarmiento, The western boundary undercurrent off the Bahamas, *J. Phys. Oceanogr.*, *16*, 233–240, 1986.
Roemmich, D., The balance of geostrophic and Ekman transports in the tropical Atlantic Ocean, *J. Phys. Oceanogr.*, *13*, 1534–1539, 1983.
Saunders, P. M., The accuracy of measurement of salinity, oxygen and temperature in the deep ocean, *J. Phys. Oceanogr.*, *16*, 189–195, 1986.
Stommel, H., and A. B. Arons, On the abyssal circulation of the world ocean, I, Stationary planetary flow patterns on a sphere, *Deep Sea Res.*, *6*, 140–154, 1960.
Swift, J. H., The circulation of the Denmark Strait and Iceland-Scotland overflow waters in the North Atlantic, *Deep Sea Res.*, *31*, 1339–1355, 1984.
Warren, B., Deep circulation of the world ocean, in *Evolution of Physical Oceanography, Scientific Surveys in Honor of Henry Stommel*, edited by B. A. Warren and C. Wunsch, pp. 6–41, MIT Press, Cambridge, Mass., 1981.
Worthington, L. V., and W. R. Wright, *North Atlantic Ocean Atlas of Potential Temperature and Salinity in the Deep Water Including Temperature, Salinity and Oxygen Profiles From the Erika Dan Cruise of 1962*, *Woods Hole Oceanogr. Inst. Atlas Ser.*, vol. 2, 24 pp., 58 plates, Woods Hole Oceanographic Institution, Woods Hole, Mass., 1970.
Wright, W. R., Northward transport of Antarctic Bottom Water in the western Atlantic Ocean, *Deep Sea Res.*, *17*, 367–371, 1970.

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