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Circulation in the eastern North Atlantic

by Peter M. Saunders¹

ABSTRACT

Basin-wide geostrophic shear profiles have been computed for several latitudes in the eastern Atlantic between 32 and 53N. The absolute velocity on each section has been chosen so that the transport satisfies the Sverdrup relation. It is concluded that the wind-driven currents are not confined to the upper thermocline but extend to great depth.

The inferred large-scale circulation south of 48N is slow (order 1 cm/s) at all levels and around 1000 m depth the meridional flow is near zero. Thus, there is very little N-S flow through the saline Mediterranean plume. Above 850 m a south-going drift develops between 48 and 30N which is fed by flow crossing the mid-Atlantic ridge from west to east principally south of 40N. Between 1200 m and 3500 m depth, water enters from the north and crosses the mid-Atlantic ridge from east to west with a flux approximately uniform with latitude. Both agreement and disagreement with other studies is noted.

1. Introduction

Leetmaa *et al.* (1977) compared geostrophic transports across selected latitudes spanning the North Atlantic Ocean with the transport predicted by the Sverdrup relation. Within the latitudes 16-32N of the anticyclonic subtropical gyre, geostrophic estimates were insensitive to the assumed level of no motion and agreed with those predicted from the curl of the wind stress. Within the cyclonic subpolar gyre, to quote the authors "there is a complicated story to tell," and between latitudes 32 and 53N no results were presented because "the Gulf Stream extends to mid-ocean in a series of loops and eddies and the data sections are dominated by transient features of large amplitude." However, east of the mid-Atlantic ridge eddying motions are much less energetic than to its west (Dickson, 1982) and since application of the Sverdrup relationship does not require full ocean-width sections, we have carried out a similar investigation employing data solely within the eastern basin. A map of the study area is shown in Figure 1 and details of the data sources are presented in Table 1.

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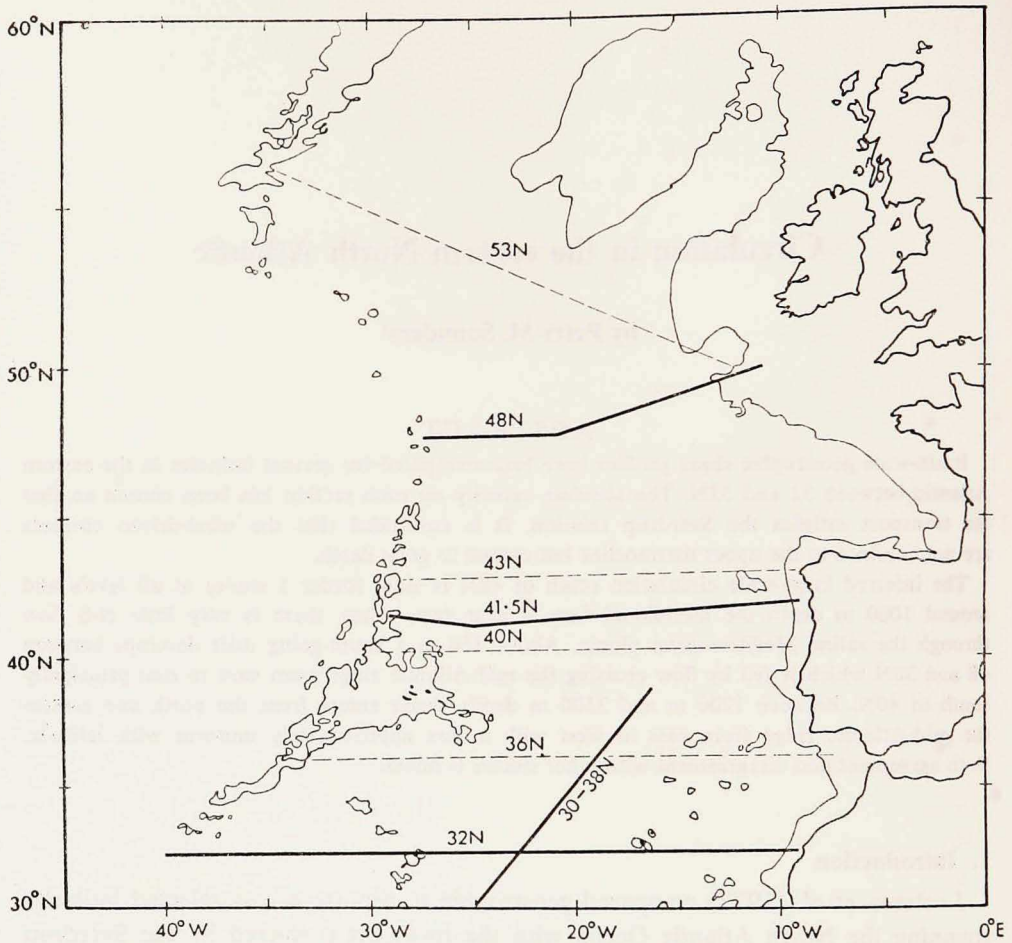


Figure 1. Map of the study area. The data sections are shown, those indicated boldly being given the greatest weight. The 1000 fathom isobath is shown.

2. The Sverdrup transport and level of no motion

Testing the Sverdrup relationship (not claimed here) requires an assumption about the level of no meridional motion which is obscure in the work of Leetmaa *et al.*, (1977). To make the point explicit the discussion begins on familiar ground. The Sverdrup transport is the sum of two parts, the Ekman transport or the frictionally determined transport of the upper boundary layer and the inviscid geostrophic transport within the water column (Stommel, 1965). The work reported here is concerned entirely with the inviscid component and, because the Ekman transport is almost always small in comparison, it is customary to speak of the Sverdrup transport (as in the heading of this paragraph) when strictly referring to the geostrophic component.

Table 1. Section data.

Latitude, °N	32	36	40	41.5	43	48	53	30-38
Longitudes, °W	10-40	8-34	10-31	10-30	9-29	11-27	11-35	17-25
Month(s), Year	11-12,57	5,59	10,57	1-2,77	9,58	4,57	8,58	1-2,81
Ship*	DI-II	CH	CF	DI	DI-II	DI-II	DI-II	DI
Stations	3631- 3649 (18)	56-73 (17)	241- 254 (14)	9329- 9406 (41)	3885- 3900 (16)	3532- 3546 (15)	3847- 3864 (18)	10270- 10294 (10)
Section length, km	2800	2300	1850	1700	1550	1150	1700	1200
Transport 10 ⁶ m ³ /s	-13 ± 3	-9 ± 2	-7 ± 2	-7 ± 2	-6 ± 2	-5 ± 2	+5 ± 2	—

* D-II *Discovery II*; DI *Discovery*; CH *Chain*; CF *Crawford*

Given geostrophy and conservation of mass one may readily derive the vortex stretching equation

$$(\rho w)_z = \frac{\beta}{f} (\rho v) \quad (1)$$

where ρ is density, f the Coriolis parameter, w the vertical velocity, v the merid-

ional velocity and β the meridional gradient of f . The suffix z denotes the vertical gradient which may be eliminated by vertical integration from a level where $W = 0$ to near the surface $W = W_E$. Thus,

$$\rho f W_E = \beta M \quad (2)$$

where M is the meridional geostrophic transport. Now $\rho W_E = \text{curl}(\tau/f)$ is the divergence of the Ekman mass flow in the near surface boundary layer and τ is the (climatological) stress exerted by the wind on the sea surface. Charts of ρW_E and M for the North Atlantic have been prepared by Leetmaa and Bunker (1978) and the estimates in this paper are derived from their calculations.

There are two problems in the application of Eq. (2) to hydrographic data. First, what is the lower limit of integration ($W = 0$) and second, what is the value of V there? Recently Killworth (1980) has shown that for geostrophic flow that conserved density a level where one component of the flow is zero is *generally* a level of absolute no motion. Thus if $W = 0$, $V = 0$ and vice versa. The following steps then provide the solution to the problem posed above: (1) choose a level of no motion, (2) calculate the transport from the surface down to that level and (3) adjust the level until the calculated transport equals that required by the wind stress data. This procedure appears equivalent to that followed by Leetmaa *et al.* (1977).

3. 1977 Section on 41.5N

The best available hydrographic data for the eastern basin was selected for study. This consisted of 41 CTD stations made from RRS *Discovery* between latitudes 41 and 42N and longitudes 10 and 30W in the winter of 1977 (Saunders, 1980). The stations were 40 km apart east-west, extended to 4000 m depth, and furnished much denser sampling of ocean-wide properties than provided by I.G.Y. data (Fuglister, 1960) which are considered later in the paper.

Matching geostrophic calculations from a pair of stations to the climatological wind stress curl makes little sense since tidal and eddy noise swamp the long term signal. Averaging over many station pairs is a procedure for reducing the noise: either historical data at fixed location may be averaged or synoptic data may be spatially averaged. The latter method which has been employed by Killworth (1980) and Schott and Stommel (1978) minimizes systematic measurement error, particularly important at depth, and seems particularly appropriate where strong boundary currents are absent. In this paper we shall adopt a basin-wide averaging scheme. On 41.5N this is unnecessarily severe, but on other sections examined later in the paper the small number of stations makes it desirable. There is no guarantee that spatial averaging of the synoptic density field will yield a close estimate of the climatological density field, but judging from the consistency of estimates made on adjacent sections the method appears to work.

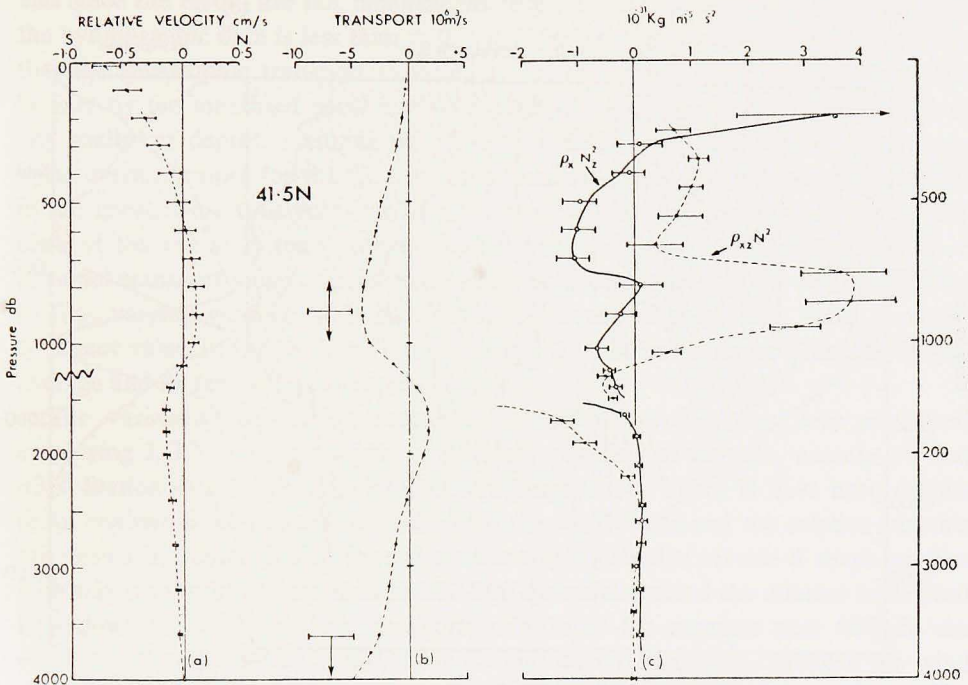


Figure 2. (a) Section average geostrophic current profile on 41.5N, 10-30W. Note change of depth scale at 1000 db. (b) Meridional transport as a function of the depth selected for zero velocity. Arrows indicate value expected from wind stress data. (c) Test of density conserving geostrophic dynamics. Theory requires equality of $\rho_x(N^2)_z$ and $\rho_{xz}N^2$ at levels where $V = 0$ but $W \neq 0$.

From tabulations of *in situ* density, level by level, one calculates the best estimate of the density gradient and its probable error, best in the least-square sense. Vertical integration of the density gradients then yields the section average geostrophic current profile².

In Figure 2a the computed profile has been referenced to 4000 db and the error estimates are shown for the shear between 4000 db and higher levels. The suitability of a linear fit of the density data is revealed by plotting the residuals from the least square fit against depth and distance, Figure 3a. There is no strong boundary-like character to the residuals which if present, would suggest alternatives to a choice of linear fit.

An eddy on the eastern flank of the mid-Atlantic ridge is the strongest feature of Figure 3a and calculations of correlations among the deviations, Figure 3b, reveal

2. At different levels a different number of stations are employed, namely 41 at depths down to 1200 db diminishing to only 8 at 4000 db. These eight deep stations are uniformly distributed across the deep part of the basin, thereby introducing no bias.

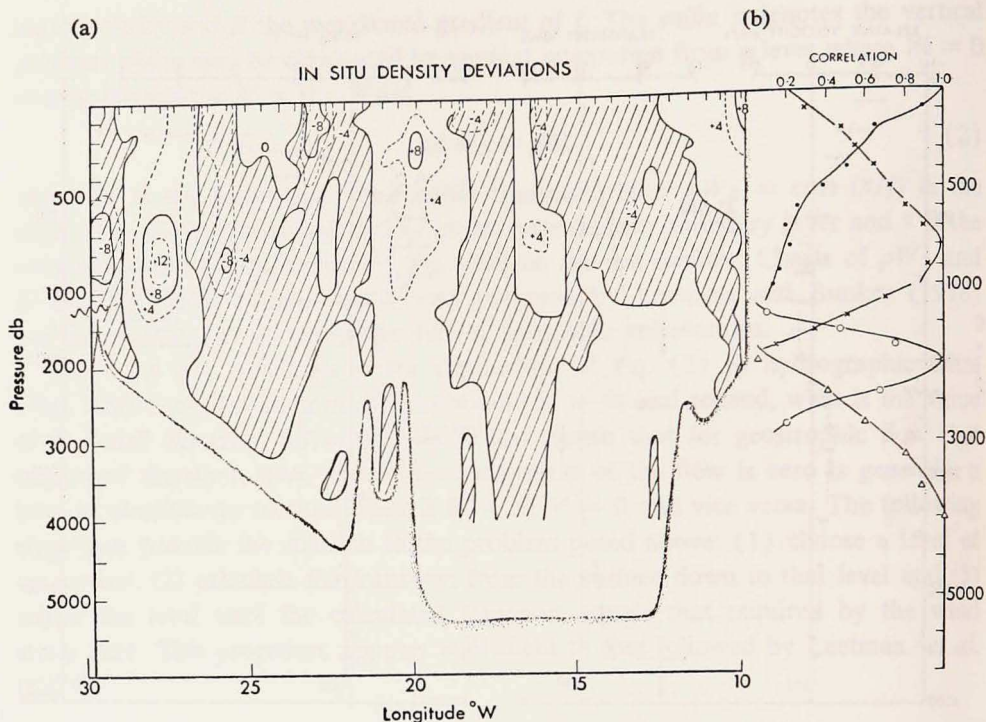


Figure 3. (a) Distribution of *in situ* density deviations from least square fit on 41.5N section (units 10^6 gm cm^{-3}). Bottom is indicated schematically. (b) Vertical correlations among density deviations. Symbol \cdot denotes correlations of surface deviations with those of an adjacent level; similarly \times for 800, \circ for 2000 and \triangle for 4000 db.

the vertical extent of it and deeper but less energetic perturbations. Note how the correlation scale increases with depth. The depth interval to a correlation of +0.5 increases from 400 m at the surface to 1500 m at 4000 m.

The geostrophic transport integrated from the surface down to the level where V and W are assumed zero is shown in Figure 2b. Because of the reverse-S shape to the current profile the transport has a maximum southward value at 900 db, reverses and has a maximum northward value at 1700 db before again becoming southward and increasing at the deepest levels on the section. (There is considerable depth in excess of 5000 db on this section but station depth was limited by winch performance.) Choice of a level of absolute no motion deeper than 900 db leads to the simultaneous vanishing of the meridional velocity at shallower levels, a situation which requires a more careful examination of both theory and data. This is deferred to the next section.

For $V = W = 0$ at 900 db the transport is $-4.0 \times 10^6 \text{ m}^3/\text{s}$, that is to the south, whereas the expected value from Leetmaa and Bunker is $-7 \pm 2 \times 10^6 \text{ m}^3/\text{s}$. Since the shear error between 900 db and the surface is at most 10% of the value

and since the errors are not independent (Fig. 3b), the transport uncertainty from the hydrographic data is less than $\pm 0.4 \times 10^6 \text{ m}^3/\text{s}$. Consequently it seems unlikely that the geostrophic transport expected from the annual average wind stress can be met by the measured geostrophic flow with a reference level at 900 db (or at any shallower depth). Leetmaa and Bunker (1978) have also calculated the seasonal average values for the Ekman divergence W_E from which seasonal variations in the geostrophic transport can be estimated. Given that the baroclinic response time of the ocean is much longer than a year (Veronis and Stommel, 1956) a seasonal transport value applied to the data is inappropriate. If the ocean response to time varying wind stress is barotropic and hence of short time scale, a winter transport value is appropriate. The value is $-2 \times 10^6 \text{ m}^3/\text{s}$ larger than the annual average and increases the discrepancy.

The variability of section averaged geostrophic profiles has been examined employing I.G.Y. data on a pair of adjacent parallels of latitude, namely 40 and 43N. Station lists supplied by NODC, Washington (see Table 1) have been subject to an analysis similar to that described for the 41.5N data and the relative currents are shown in Figure 4. All three show remarkably similar reverse-S shape profiles with relative northward maxima at 800-900 db which exceed the relative northward flow down to 4000 db. The transports relative to the maxima near 900 db are -2.8×10^6 and $-3.4 \times 10^6 \text{ m}^3/\text{s}$ for 43 and 40N respectively, whereas the wind stress estimates are $-6 \pm 2 \times 10^6$ and $-7 \pm 2 \times 10^6 \text{ m}^3/\text{s}$. The three data sets are separated not only by season but also by two decades and exemplify the remarkable stability of the density field in striking contrast to the unsteadiness of direct current measurements. The two I.G.Y. sections also confirm the discrepancy between the transports deduced from the wind and hydrographic data sources found for the 41.5N section.

4. The depth of the Sverdrup transport

The error in the estimate of curl (τ/f) is reckoned by Leetmaa and Bunker (1978) to be $\pm 30\%$. If the wind stress was reduced systematically by this amount, wind derived transports would more nearly agree with the hydrographic estimates for latitudes 40-43N. There is, however, no evidence for such a discrepancy in the results of Leetmaa *et al.* (1977). Alternatively, the possibility must be considered that the wind driven transport is not confined to the upper ocean.

If a level between 1000 and 4000 db is chosen as a level of absolute no motion, then on all three sections between 40 and 43N the meridional velocity V (but not W) simultaneously vanishes at shallower depths. For example on 41.5N, (see Fig. 2b) the geostrophic transport has the value $-7 \times 10^6 \text{ m}^3/\text{s}$ required by the wind stress only by extrapolating the measured shear to 4300 db³.

3. The zero of V is then $+0.25 \text{ mm/s}$ in Figure 2a, a barely perceptible shift.

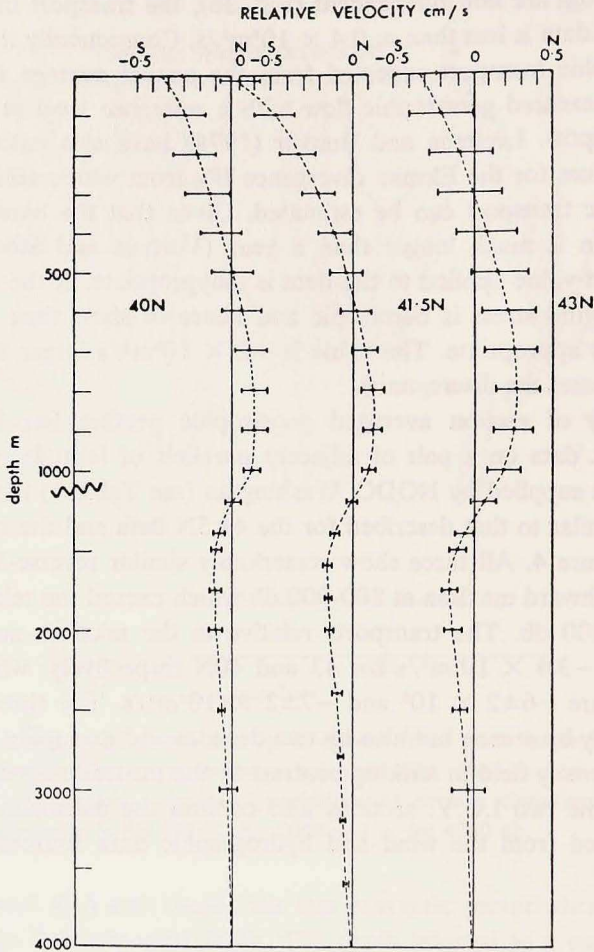


Figure 4. Section average geostrophic current profiles on adjacent parallels spanning a 20 year interval.

At 4300 db $V = W = 0$ at 1200 and 600 db $V = 0$ but $W \neq 0$. The rule described earlier appears violated. Killworth (1980) shows that this situation may arise if Jacobian-like functions of the density field are zero. The condition (in section 2 of his paper) may be written

$$\rho_x(N^2)_z - \rho_{xz} N^2 = 0$$

where N is the Brunt-Väisälä frequency, a measure of the vertical density gradient taking into account compressibility. On 41.5N the quality and quantity of CTD data allow estimates of the above quantities to be made at 100 m intervals (shallow) increasing to 400 m intervals (at depth): $\rho_x(N^2)_z$ and $\rho_{xz} N^2$ are plotted individually

in Figure 2c. Near 1200 db these quantities *are* equal as required by the theory but at 600 db they are *not*. Both assertions depend critically on error estimates which are shown on the same figure. At a given level N^2 averaged over the section has a simply calculated standard error and the vertical gradient is formed by differencing values at adjacent levels and calculating its error by assuming the two error estimates independent. Similarly, the least square density gradient at a particular level has a well-defined error measure and the vertical gradient ρ_{xz} and its associated error are determined in like manner. No allowance is made for the finite difference approximation to the vertical gradient in either case. If the observations fail to satisfy the requirements of the theory, as appears to be the case, the most probable conclusion is that the large scale mean motions of the ocean do not conserve density. A recent paper by Keffer and Niiler (1982) arrives at this conclusion and if generally true, one can no longer *assert* that $V = 0$ where $W = 0$ one must *assume* it so.

The vanishing of V and W at great depth found on 41.5N is in accord with the early views of Stommel (1965, p. 154) but at variance with recent claims that the wind driven circulation occupies only (circa) the upper 1000 m of the ocean (Leetmaa *et al.*, 1977). A shallow layer is also found in the numerical modelling of Holland and Rhines (1980). The writer finds a deep level of no motion plausible for the following reason. At 4000 m salinity in the eastern basin is a remarkably exact function of potential temperature (Worthington and Metcalf, 1961; Saunders, unpublished) suggesting that sources and sinks are unimportant in relation to isopycnal stirring. A number of small interconnecting passages into the western basin exist deeper than 4000 m, all at very low latitudes; otherwise the water at 4000 m is entirely confined. Basin-wide average flow must be small in these circumstances. In contrast to this at latitudes north of 50N, sources of deep water and sinks of surface water in the Norwegian sea permit large basin average meridional velocities at somewhat shallower depths, as is apparent later. This thermohaline flow is accompanied by large variations in the relationship between salinity and potential temperature.

5. Other I.G.Y. sections

In Figure 5 are found section average geostrophic current profiles (referenced to 4000 m) for latitudes 32, 36 and 48N with transport calculations as a function of the choice of level of absolute no motion in Figure 6. The entire 32N section was discussed by Leetmaa *et al.* (1977). Here, only the half east of the mid-Atlantic ridge (at 40W) is considered (see Table 1). The similarity of the current profiles on 32 and 41.5N is evident. A meridional velocity of zero at 1000 m where $W = 0$ yields a transport of $-13 \times 10^6 \text{m}^3/\text{s}$ as required by the wind data. Thus, the wind-driven layer for 32N is shallow as reported by Leetmaa *et al.*, (1977).

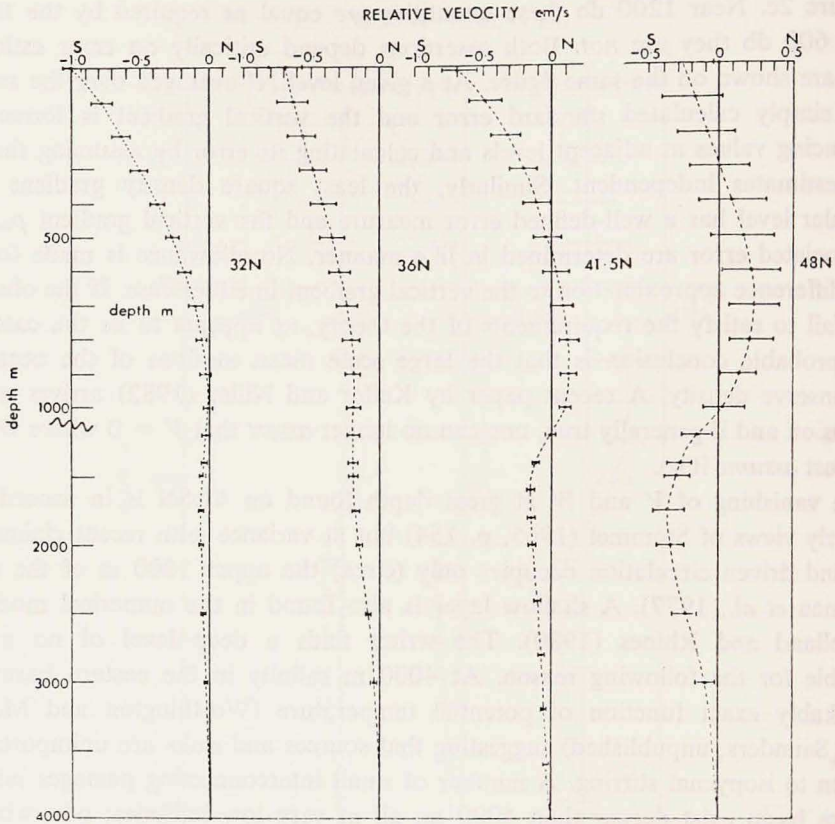


Figure 5. Section average geostrophic current profiles for latitudes 32-48N: the meridional velocity is assumed zero at 4000 m.

However, a deep level of absolute no motion at 3200 m is just satisfactory and cannot be distinguished from the shallow given the errors in computed and geostrophic transports. (A fact also recognized by Leetmaa *et al.*, 1977.) On the 36N section the current profile lacks the reversal of shear above 1500 m and requires $V = W = 0$ at 2700 m to give the required $9 \times 10^6 \text{m}^3/\text{s}$ transport to the south. The section appears inconsistent with other current and transport profiles (Figs. 5, 6) and will be given no further consideration here. The section has recently been repeated and the new data is awaited with interest.

The 48N section has a highly developed reverse-S shaped profile with northward shear extending above 1500 m to shallow depths thereby permitting only a weak surface southward flow. The transport of $-5 \times 10^6 \text{m}^3/\text{s}$ required zero meridional velocity at 4000 m where $W = 0$ and gives simultaneously $V = 0$ at 300 and 1000 m.

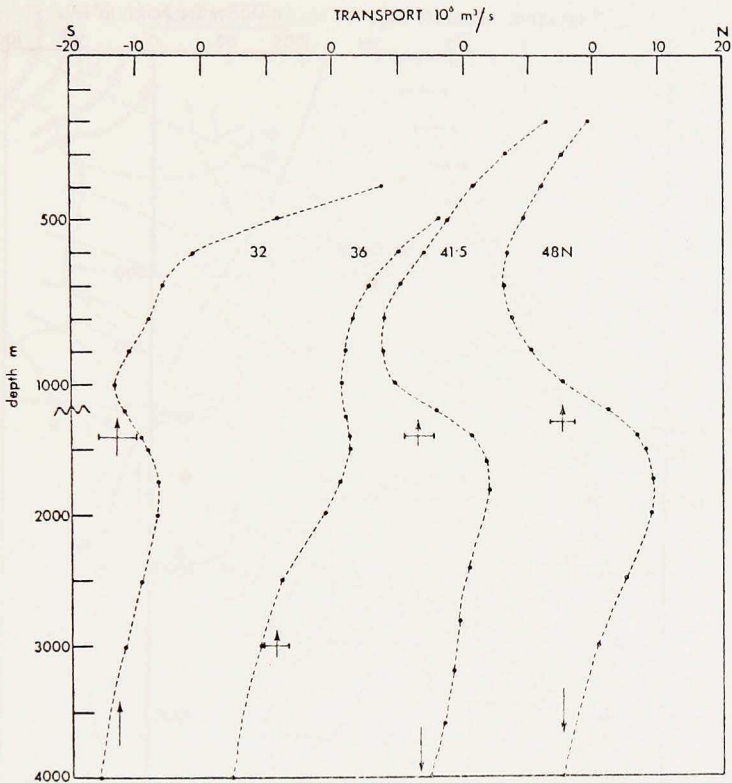


Figure 6. Meridional transport integrated from the top to the bottom of the ocean as a function of the depth selected for zero velocity. Arrow indicates expected value from wind stress data. Each curve is displaced $+20 \times 10^6 \text{ m}^3/\text{s}$.

The sections considered above all reside within the anticyclonic subtropical gyre. A section on 53N lies principally within the subpolar gyre and has somewhat different properties. The section average geostrophic current profile, Figure 7a, has northward shear at all levels above 2500 m and currents and transports, Figure 7b, are three times larger than on the sections to the south. Zero meridional velocity at 1100 m yields a net transport of $+5 \times 10^6 \text{ m}^3/\text{s}$ as required by the wind data, made up of $+20 \times 10^6 \text{ m}^3/\text{s}$ between the surface and 1100 m and $-15 \times 10^6 \text{ m}^3/\text{s}$ below this level. On this section the transport is integrated from 0 to 4000 m. At the bottom $W = 0$ but $V = 0$ leads to a northward transport of approximately $30 \times 10^6 \text{ m}^3/\text{s}$. A mid-depth zero of meridional velocity approximately balances northward and southward transports as required by the wind stress curl but yields no level of absolute no motion. The difference between the earlier sections and that on 53N undoubtedly resides in the strength of the meridional thermohaline circulation on the latter. Yet it is noteworthy that the

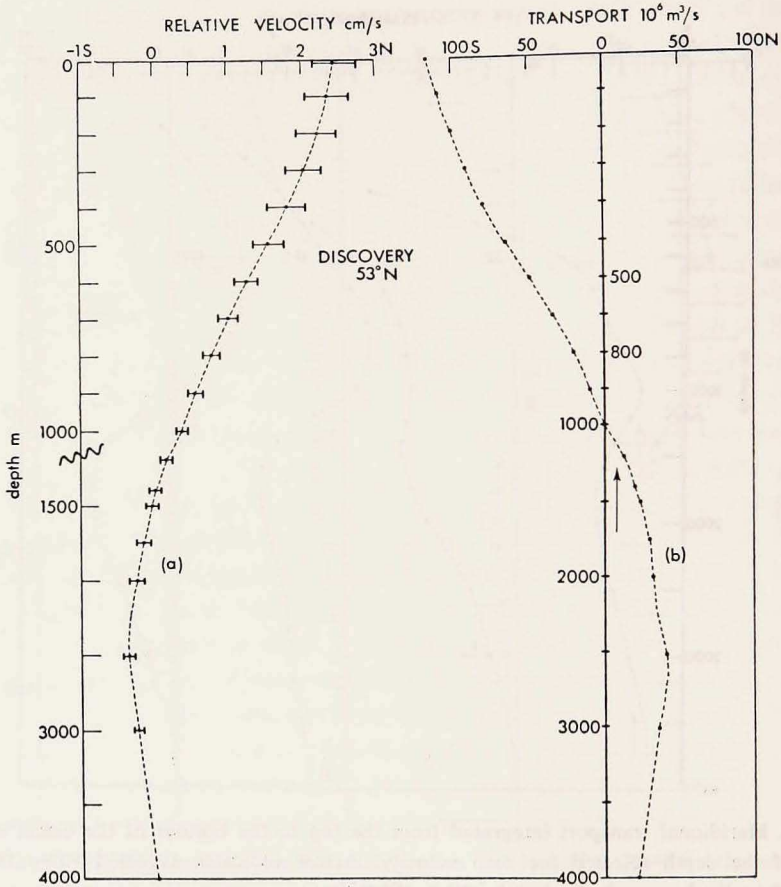


Figure 7. (a) Section average geostrophic current profile on 53N, I.G.Y. Data. (b) Transport normal to section integrated from the top to the bottom of the ocean as a function of the depth selected for zero velocity. Arrow indicates value expected from wind stress data.

meridional flow is very weak at 1000 ± 100 m throughout the entire region studied in this paper.

6. The eastern Atlantic between 30 and 50N

The deep level of no motion within the subtropical gyre implies that the differences between absolute and 4000 m relative profiles are very small and for illustrative purposes (only) can be ignored. An examination of these profiles (Fig. 5) and a knowledge of the Mediterranean outflow suggests that the main features of the circulation in the eastern North Atlantic can be described within the framework of a three layer structure; an upper layer 0-850 m, a deep layer 1200-3500 m and an intermediate layer from 850-1200 m, the last encompassing the spreading

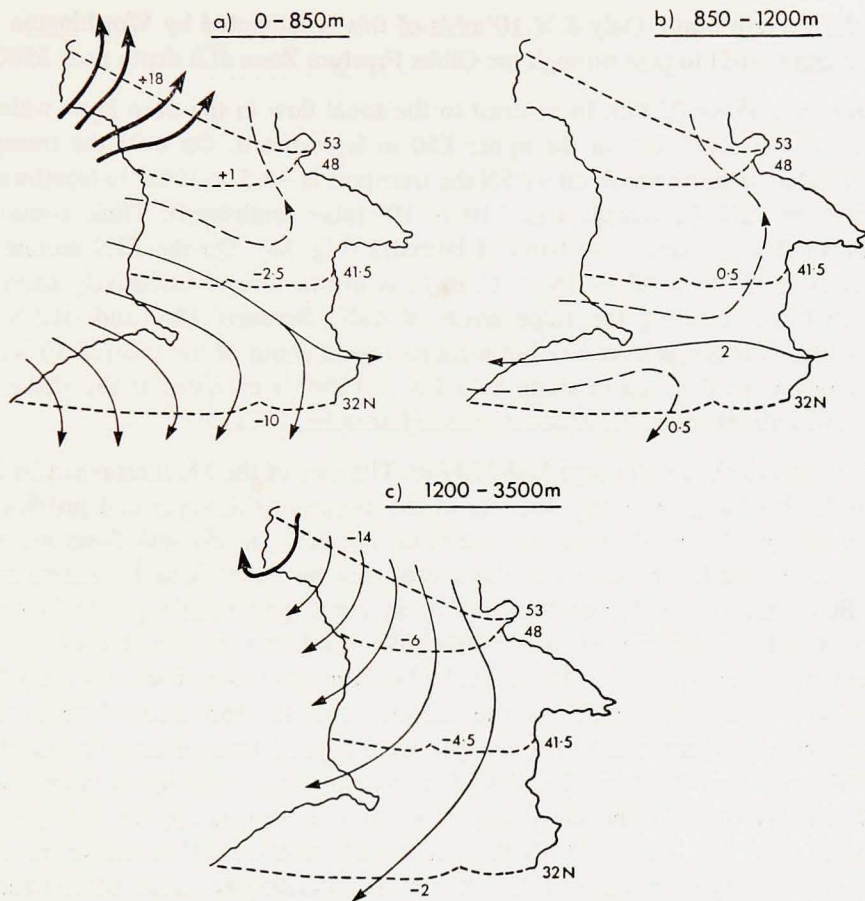


Figure 8. Basin wide circulation of eastern North Atlantic (schematic) (a) Depth range 0-850 m, (b) 850-1200 m, (c) 1200-3500 m. Transport in units of $10^6 \text{m}^3/\text{s}$.

saline Mediterranean water. The deduced transports are shown schematically in Figure 8 where variations in calculated meridional transport are assumed met by flow across the mid-Atlantic ridge. Some corroboratory evidence for the inferred zonal transport is presented later.

a. Transport below 1200 m. Southward transport is present on all sections and decreases with latitude. Consequently deep water leaves the eastern basin crossing the ridge from east to west (Fig. 8c) although this outflow cannot extend much below 3000 m because of the height of the ridge crest. Across 48, 41.5 and 32N the (southward) transports are -6 , -4.5 , $-2 \times 10^6 \text{m}^3/\text{s}$ respectively, showing an approximately uniform cross-ridge flow with latitude. Between 53 and 48N there is a large change, from -14 to $-6 \times 10^6 \text{m}^3/\text{s}$, implying an outflow of $8 \times 10^6 \text{m}^3/\text{s}$

into the western basin. Only $5 \times 10^6 \text{ m}^3/\text{s}$ of this is estimated by Worthington and Volkmann (1965) to pass through the Gibbs Fracture Zone at a depth near 3500 m.

b. Transport above 850 m. In contrast to the zonal flow in the deep layer which is westward, the zonal flow in the upper 850 m is eastward. On 48N the transport is $1 \times 10^6 \text{ m}^3/\text{s}$ (northward), on 41.5N the transport is $-2.5 \times 10^6 \text{ m}^3/\text{s}$ (southward), whereas on 32N the transport is -10×10^6 (also southward). Thus, continuity requires inflow for the entire band of latitudes (Fig. 8a). On the 53N section the transport in the layer of $+18 \times 10^6 \text{ m}^3/\text{s}$ is drawn almost exclusively from the western basin, crossing the ridge north of 48N. Between 48N and 41.5N the cross-ridge flow is much weaker but increases again south of the Azores. From the southern half of the area of study 1 to $1.5 \times 10^6 \text{ m}^3/\text{s}$ of water at the shallowest depths (0-200 m) enters the Mediterranean (Lacombe, 1971).

c. Transport in the depth range 850-1200 m. The axis of the Mediterranean outflow has a depth of approximately 1000 m at the center of this layer and justifies the choice of boundaries. As may be seen from Figure 5, meridional flows are very small within the layer and zonal flows can only be determined by examining a meridional section. In Figure 9 a section averaged geostrophic profile is shown from 10 CTD stations made between 30N 25W and 38N 17W from RRS *Discovery* in January-February 1981 (Fig. 1; Table 1). Consistent with finding a deep level where $W = V = 0$ we choose the deepest level for the zero of zonal flow. (Quantitative but not qualitative changes would result from taking any level up to 3000 m.) Near 1000 m the 1000 km average zonal velocity is estimated as -0.7 cm/s (westward) and the transport within the layer is approximately $-2 \times 10^6 \text{ m}^3/\text{s}$ (Fig. 8b). Across 41.5N there is a weak northward flow drawn from the western basin: by entrainment this could assist spreading of saline Mediterranean water along the European coast at depths near 1000 m.

The zonal flows within the shallow and deep layers inferred from conservation of mass are clearly corroborated by the geostrophic profile in Figure 9. At depths less than 750 m the flow has an eastward component and at depths greater than 1200 m the flow has a westward component. Thus a consistent picture of the circulation emerges from the choice of a level of absolute no motion which satisfies the requirements of the wind stress curl.

7. Comparison with other studies

Many distinguished oceanographers have studied the circulation of the North Atlantic. A comprehensive review of the mid-depth circulation which also touches on near surface flow has been given by Reid (1981). A 110 page monograph by Worthington (1976) also summarizes earlier studies. Both papers have an enormously wider scope than has been attempted here and space permits us only to mention findings which illuminate or contrast with their work.

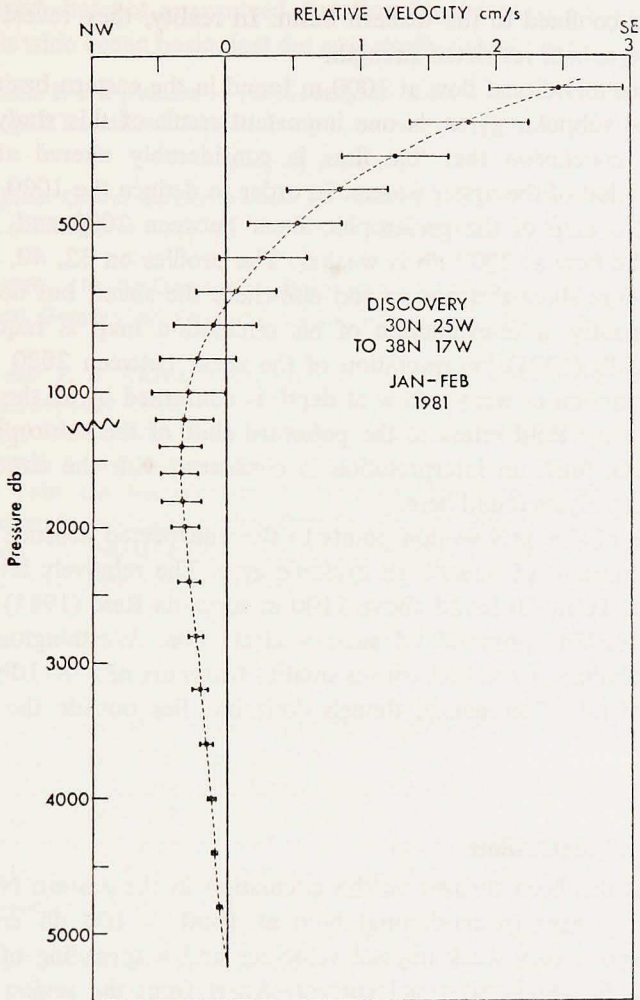


Figure 9. Section average geostrophic current profile on deep axis of eastern basin between latitudes 30 and 38N. The upper flow is SE, the deep flow, taking account of Figure 5, is SW.

Worthington was much concerned with the 'Gulf Stream recirculation' and proposed a small intense anticyclonic gyre in the western basin, arguing that an Atlantic-wide gyre would sweep aside the Mediterranean outflow. This work confirms Worthington's view that the upper circulation in the eastern basin is weak but finds an eastward extension of anticyclonic gyre from the surface down to a depth of 850 m. Below 1200 m (in Worthington's class 4-7°C) the flow is again anticyclonic and perhaps represents an eastward extension of his 'northern gyre.' It is too easy to interpret Worthington's circulation diagrams as revealing

that the flow is confined to the western basin. In reality, they reveal only that an Atlantic-wide gyre is of restricted strength.

The near zero meridional flow at 1000 m found in the eastern basin, both in the subtropical and subpolar gyres, is one important result of this study. It confirms Reid's (1978) conclusion that 'the flow is considerably altered at mid-depths' compared with that of the upper waters. In order to deduce the 1000 m circulation Reid prepared a map of the geostrophic shear between 2000 and 1000 db and assumes that the flow at 2000 db is weaker. The profiles on 32, 40, 41.5, 43, 48, 53N deduced here show this not so and elsewhere the shear, but not the flow, is weak. Consequently, a re-evaluation of his circulation map is required. This is not true of Reid's (1981) interpretation of the shear between 2000 and 3500 db where the assumption of weaker flow at depth is confirmed by all the profiles here. In this depth range Reid refers to the poleward shift of the subtropical gyre with increasing depth. Such an interpretation is consistent with the circulation in the shallow and deep layers found here.

The analysis of the 53N section points to the unexpected absence in Worthington's circulation maps of a northern *cyclonic* gyre. The relatively large northward transport ($20 \times 10^6 \text{m}^3/\text{s}$) found above 1100 m supports Reid (1981) and Leetmaa and Bunker's (1978) proposal of such a circulation. Worthington has only a vertical thermohaline circulation with a smaller transport of $8 \times 10^6 \text{m}^3/\text{s}$. Further investigation of this discrepancy, though desirable, lies outside the scope of the present paper.

8. Summary and conclusions

Certain light has been thrown on the circulation in the eastern North Atlantic. The choice of a zero of meridional flow at 1000 ± 100 db ensures correct Sverdrup transport, very weak abyssal velocities and a spreading of the Mediterranean outflow in an almost zonal current. Apart from the region of the saline plume, continuity then imposes nearly zero zonal flow at 1000 db. Thus, conventional maps of the 0-1000 db dynamic height anomaly, say east of 40W might well be regarded as maps of the upper circulation of the ocean (see for example, Reid, 1978). Yet most readers will have observed that many height contours intersect the shelf and have been puzzled by it. Ageostrophic currents in the surface mixed layer (Ekman flow) cause fluid to cross height contours but this flow is restricted to the upper boundary layer. Holland and Rhines (1980) in an analysis of a 'realistic' numerical general circulation experiment emphasize that eddy stresses associated with an enstrophy cascade produce the same effect through the depth of the fluid. If eddy stresses are important, the slow mean circulation of the ocean is not geostrophically controlled and the analysis presented in this paper (and numerous others besides) is of doubtful value. Reassuringly, the same authors find

that the preferred, but not guaranteed, location for Sverdrup dynamics is in the eastern half of a wide ocean basin, just the area studied in this paper.

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