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Inversions of observations near the Azores Front

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ABSTRACT

CTD data from *Discovery* Cruise 138 of late June, 1983 in the Madeira Abyssal Plain show a strong, but isolated, Mediterranean Water (MW) signal around 35N, 25W. This water mass gives rise to an anticyclonic eddy at its southward tip and, through double diffusive processes, to modification of the density field a few hundred meters below. Current estimates have been calculated from the cruise data using the Bernoulli inverse technique. Extensions of this inverse method, which take into account the depth dependence of dynamically informative density and potential vorticity functions, are developed to investigate the flow field and the consistency of the inversions. The Bernoulli inversions all show a strong eastward current in the upper 500 m around 34–35N. In the south of our study area the upper level flow is predominantly southward, consistent with current meter records from the NEADS 1 site at 33N, 22W. It is speculated that during the summer of 1983 there was a southward excursion of the Azores Front into this area, as has been recorded in the past.

1. Introduction

Since the classical view given by Wust (1935) of the upper level flow in the North Atlantic the region between the Azores and Madeira has been seen as the location of significant gradients in the direction of the mean flow, associated with the northeast corner of the subtropical gyre. More recent work using dynamic height calculations from historical data (Pollard and Pu, 1985; Maillard, 1986) confirm the picture of the upper level flow in this region altering from predominantly eastward in direction near the Azores to predominantly southward near 20W (see Fig. 1). A zone of rapidly changing properties has been observed a number of times in this area (Kase and

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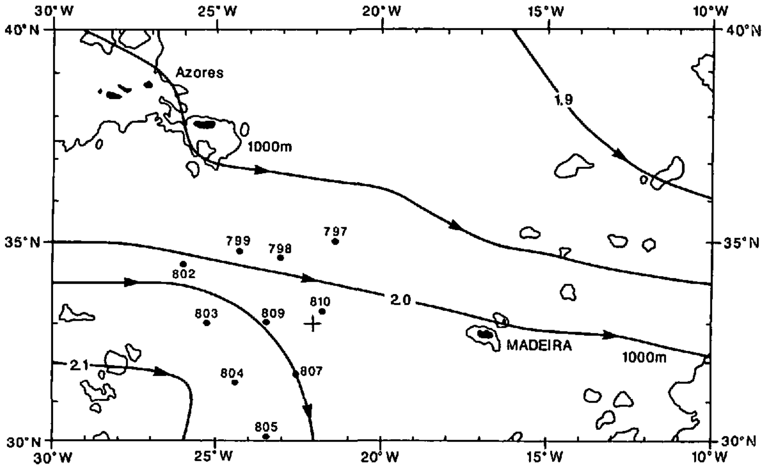


Figure 1. Location of *Discovery* Cruise 138 stations used in this paper, and the NEADS 1 current meter (shown by a cross). The annual mean dynamic height at 50 db, relative to 3000 db, is also shown (after Maillard, 1986).

Seidler, 1983; Kase *et al*, 1985; Pollard and Pu, 1985) associated with the anticyclonic curvature of the flow; it is called the Azores Front.

Discovery Cruise 138 made a CTD survey of the region in late June–early July of 1983. The locations of the stations are shown in Figure 1. Data from this cruise have been previously reported in Gould (1983), Saunders (1987) and Saunders and Dolan (1987). However, the 10 stations in the Madeira abyssal plain have not been fully analyzed. As they are in the region where the Azores Front has been observed the stations offer an opportunity to investigate this dynamical feature further. Also, as they lie near the site of the NEADS 1 current meter mooring (Muller, 1984) the stations provide an ideal site for the validation of inverse techniques for finding ocean circulation.

This paper first examines property distributions over the core of the observed area. The potential temperature, salinity and sigma-*n* distributions all show evidence of a front in the north of the study area curving to the south on the eastern edge in the upper few hundred meters. In addition, a strong Mediterranean water intrusion, or possibly a meddy, centered on station 799 (34° 40′N, 24° 20′W) at a depth of 700–1300 m, produces evidence for the existence of a deep eddy. To support inferences drawn from these distributions the Bernoulli (Welander, 1983, and particularly Killworth, 1986) inverse method is applied to the data in order to reconstruct the flow fields. It produces a velocity field broadly consistent both with the data and current meter measurements available for the same period. Numerical experimentation with the form of the density variable used in the Bernoulli method yields a consistent image of the direction and magnitude of the current in the study region, and its vertical shear.

The results show that the Bernoulli method can be used, with care, on semi-synoptic data whose time variability is on a scale only slightly longer than the period taken to make the measurements.

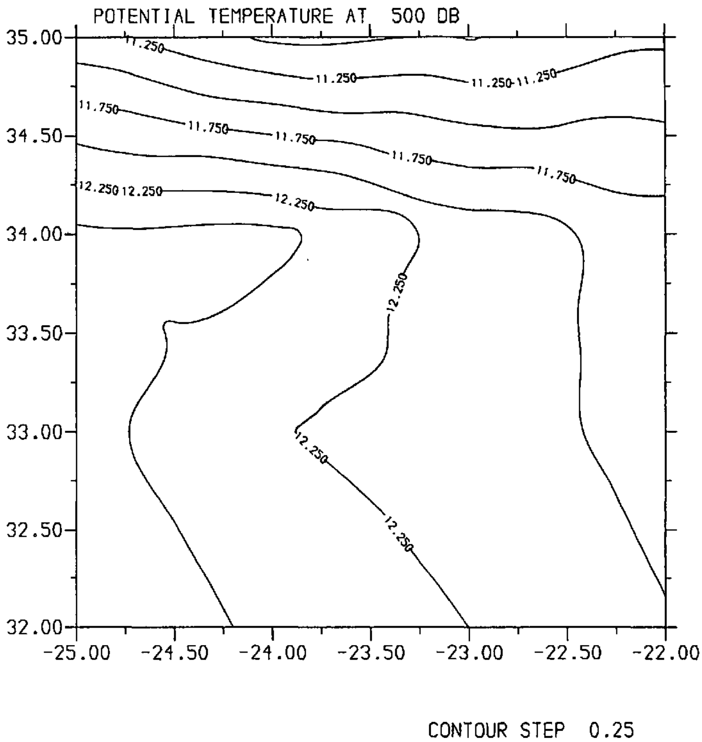
2. Data presentation and analysis

The raw CTD data for the 10 stations illustrated in Figure 1 have been presented in Saunders and Dolan (1987). The most striking feature of the vertical station profiles shown there is the strong MW intrusion between 700 m and 1300 m depth at station 799 (34° 40′N, 24° 20′W). The salinity reaches 36.1 and the temperature rises to 12°C near 1000 m. Some of the other stations in the set show evidence for MW influence in the same depth range, but only very localized zones reach salinities of 35.7 or temperatures of 9°C.

In Figures 2–4 the potential temperature, salinity and sigma-*n* distributions respectively are shown for two pressure surfaces within the core of the study area. For all three properties the uppermost pressure levels (e.g. 500 db) have a strongly zonal distribution in the north of the study area, consistent with, although suggesting stronger gradients than, climatology (Robinson *et al.*, 1979; Levitus, 1982; Maillard, 1986). The thermal wind balance indicates the presence of a strong eastward current, with large vertical shear. However, south of 34N the distributions tend to become meridional in character, consistent with a southward current.

At 1000 db the MW intrusion from the north becomes the dominant feature. The salinity and temperature distributions both show a tongue of warm, salty water extending from 35N, 24W into the region. The sigma-*n* distribution has an anticyclonic closed contour feature at the tip of this intrusion (see Fig. 4b). Such a light density structure associated with an MW tongue was reported by Arhan and de Verdiere (1985) in the Tourbillon region, and was ascribed to double diffusive convection. The vertical sections in Saunders and Dolan (1987) show considerable structure in the MW signal below 1000 m supporting the postulated presence of double diffusion. The MW signal is still evident at 1600 db (not shown), where warm, salty water is present beneath the tip of the MW intrusion (providing more support for double diffusive convection) with an associated low density. By 2000 db a large region of weak gradients has been established, except in a narrow zone across the middle of the area. A tendency for zonality becomes more pronounced with depth, with very weak gradients becoming the norm.

Compared with the climatological distribution of these properties the data from this cruise would seem to be anomalous, and therefore possibly strongly time varying. The stations were occupied over a period of 10 days, which in oceanic terms should be roughly synoptic. We are fortunate in having the observations from a long-term current meter mooring adjacent to our study area with which to check the temporal gradients. This is the NEADS 1 site (Muller, 1984) at 33° 10′N, 21° 55′W, close to station 810 (see Fig. 1). The relevant deployment ran from 19 April–19 October 1983 and there were six sets of current meter, temperature, salinity and pressure sensors



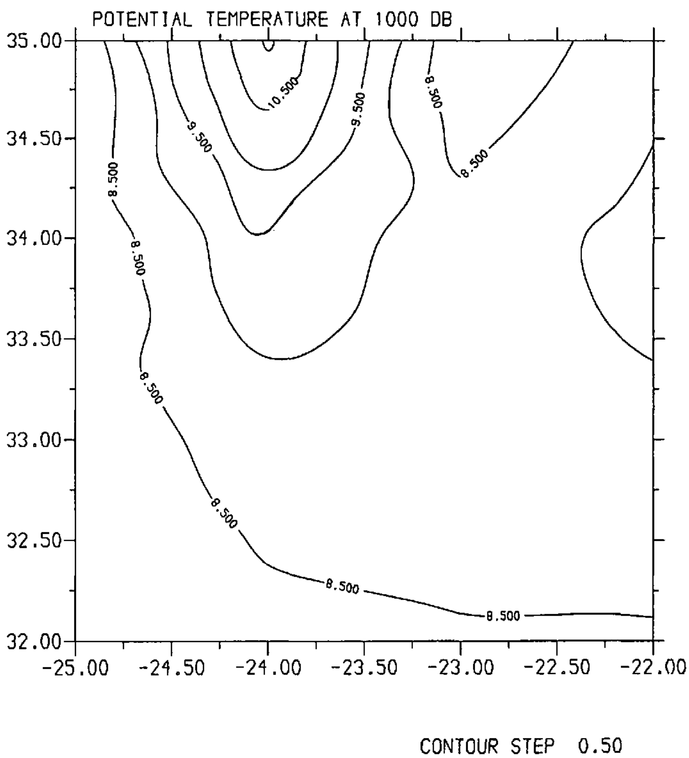


Figure 2. Potential temperature distribution at two levels in the core of the study area.

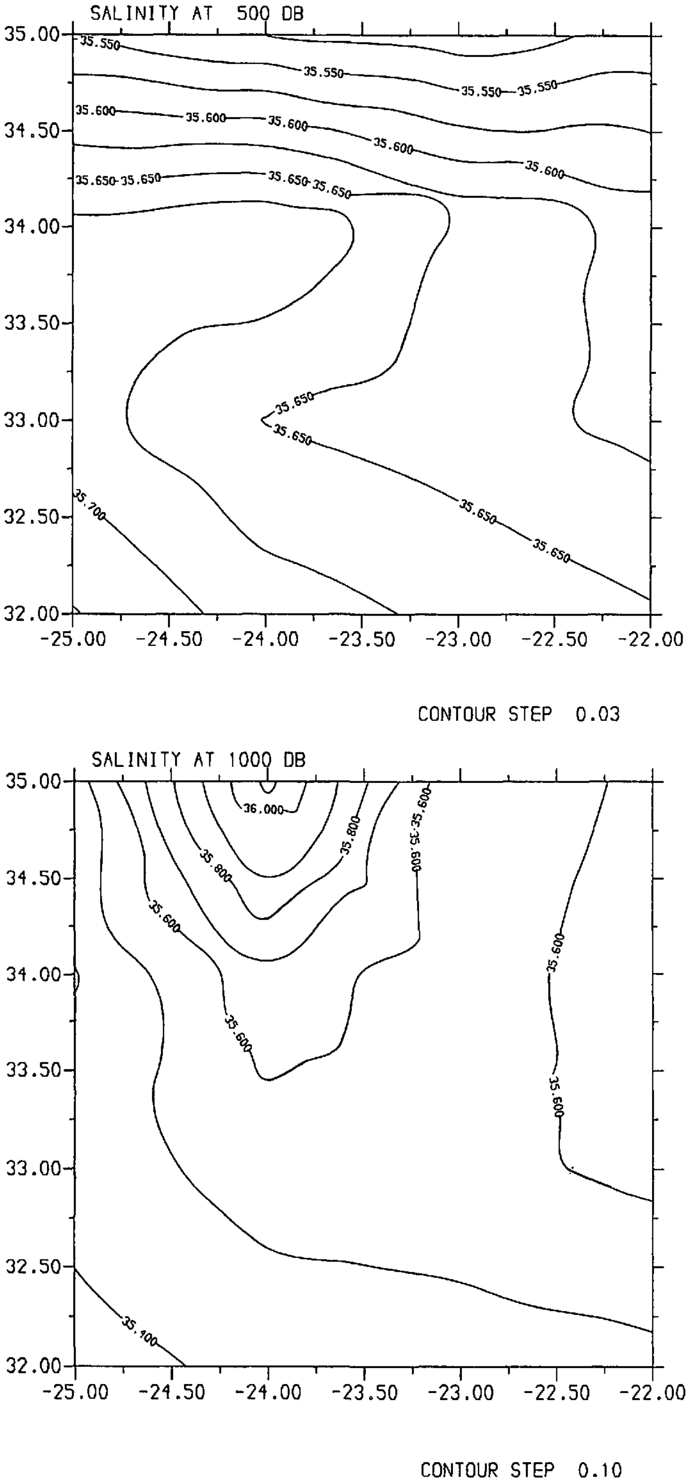


Figure 3. Salinity distribution at two levels in the core of the study area.

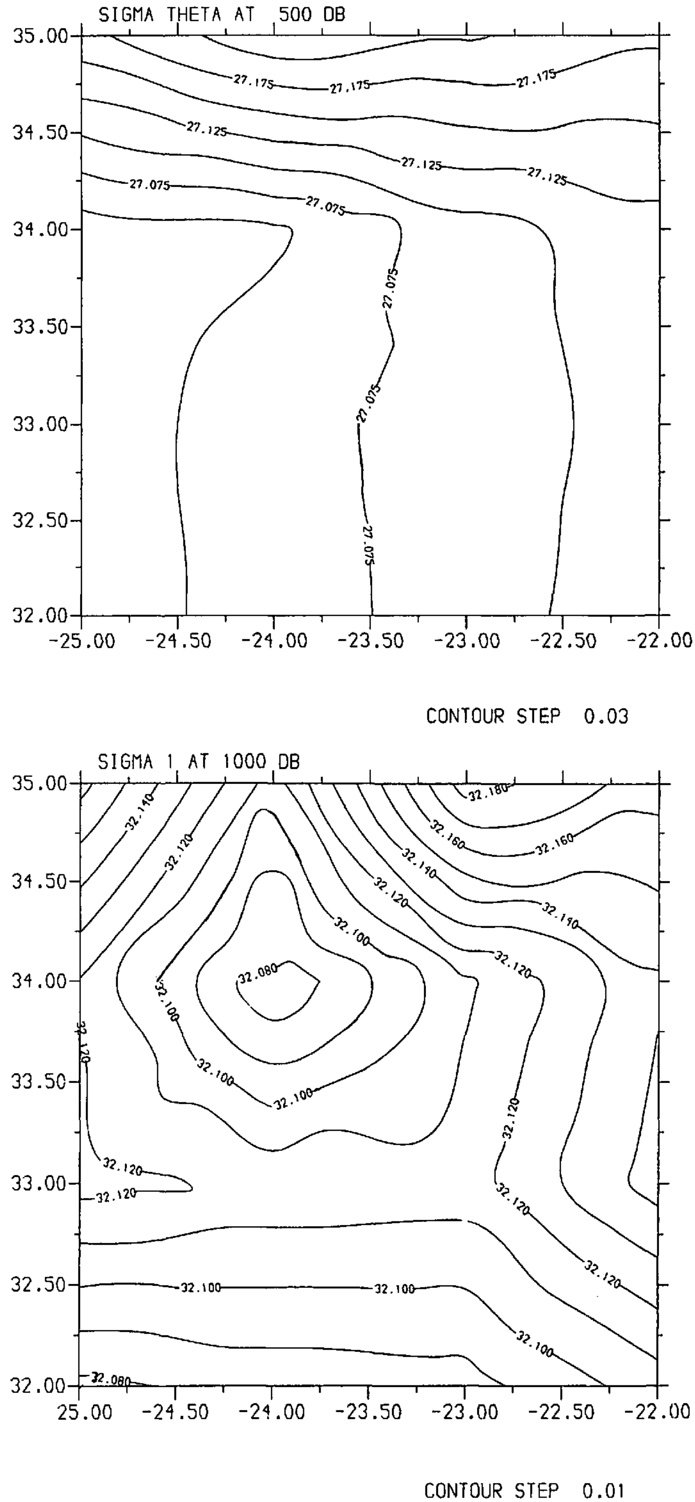


Figure 4. Sigma-*n* distributions at two levels in the core of the study area.

deployed at depths ranging from 245 m to 2980 m. Despite the nearby presence of the MW intrusion all the records have relatively weak temporal gradients during June and July (see Fig. 8). The current meters also confirm the presence of a strong and persistent southward flowing current near 22W, which has a pronounced vertical shear between 500 m and 1500 m.

3. Description of the inverse technique

Given the apparently temporally stable flow field suggested by the NEADS 1 mooring, the data provide a good test-bed for inverse calculations of the velocity field, as one of their main assumptions is steady flow. The Bernoulli inverse method was used to study the region, because of the robustness of its behavior over widely differing regions of ocean in the sensitivity analysis of Killworth and Bigg (1988). A short resume of the method follows:

This inverse technique was developed by Killworth (1986), after some earlier work of a similar nature by Welander (1983). Assuming a steady-state ocean that is in approximate geostrophic and hydrostatic balance, the method conserves mass exactly and density approximately. Linear potential vorticity $q = f\rho_z$ is therefore also approximately conserved (Welander, 1971), where f is the Coriolis parameter, ρ the density, and z a coordinate vertically upwards. A third scalar, the Bernoulli function $B(=P + \rho gz)$ is also conserved approximately, where P represents the pressure and g the acceleration due to gravity. Under most conditions this tripartite conservation implies that one scalar is a function of the other two, for example, $B = F(\rho,q)$. If the vertical profile of density ρ is known at a station, then q is calculable and B also, from the hydrostatic relation, to within an unknown additive surface value.

To determine this unknown constant the method examines a collection of stations pairwise, seeking for a depth at the first station at which ρ and q match the values at some other depth at the second station (the search begins below the mixed layer so that q is well defined). Forcing B to match also gives a linear equation connecting the difference between the surface pressures at the two stations. Typcially there are multiple such matchings for any pair of profiles so that the full system is heavily over-determined and can be solved by singular-value decomposition in a least-squares sense. Velocities can then be computed from numerical differentiation of the pressure field.

This method does not need to explicitly compute horizontal gradients until the basic pressure field has been constructed from finding common streamlines of the flow. Most inverse techniques need such gradients as an integral part of their calculations, thus introducing significant errors into the methods due to the difficulty in evaluating these gradients accurately (Bigg and Killworth, 1988). However, in a frontal region, such as is encountered in this data, a probable transition from one functional relationship between B, ρ and q to another is likely to occur. This is suggested for near surface waters in Figure 5, which shows a plot of σ_θ versus q for the depth profile at each

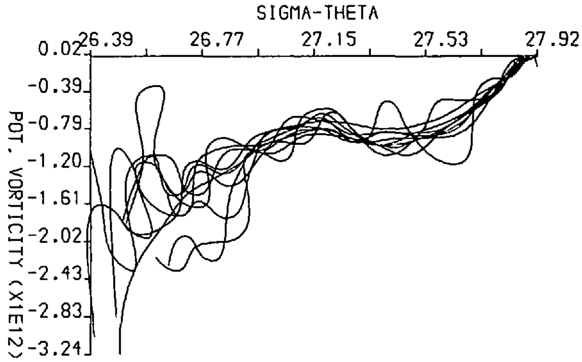


Figure 5. Sigma-theta versus potential vorticity with depth for the 10 stations.

station. Note, though, that for much of the water column, below $\sigma_\theta = 26.9$, the profiles follow a similar form. It may therefore be expected that upper level inversions in the vicinity of the front may be distorted. Comparison with the NEADS-1 records, and a general consistency check with the observations are necessary to assess such distortion. Another source of noise is in the searching for matchings of ρ and q, as q is derived from a cubic spline fit to individual $\partial\rho/\partial z$ values. It should be stressed that the Bernoulli method has not been previously applied to synoptic data, as other inverse methods have, so this paper is testing its reliability.

In common with some other inverse methods, it is not clear what density variable is appropriate for the method. It has been cogently argued that surfaces of σ_θ do not necessarily represent flow surfaces at depth (Long, 1985; McDougall, 1987) and analyses of full depth data using depth-varying density variables give a more representative picture of the deep flow fields (Harvey and Arhan, 1988).

The dependence of the Bernoullean velocity field on the density variable was tested in several ways. Inversions were performed with σ_θ as the variable throughout the water column. Others carried out the pairwise matching over limited depth ranges using locally referenced potential densities. A third technique was the development of a Bernoulli-type method where the potential density variable used for matching was allowed to vary with depth (eg. σ_θ from the bottom of the mixed layer to 500 m, σ_1 from 500–1500 m, σ_2 from 1500–2500 m, etc). This choice limited the pairwise matching to the particular variable layers, but as the number of matchings is of the order of 600 it was not thought that very much information was lost by this restriction. Indeed, the number of matchings actually *increased,* in comparison to the sigma-theta trial, from 648 to 788. Some extra constraints had also to be added to the final set of equations to ensure that the Bernoulli function was continuous across the layer interfaces. From Killworth (1986; Eq. 11)

$$B_n = B_n(\text{surface}) + g\int z(\partial\rho/\partial z)\,dz \qquad (1)$$

so that for the Bernoulli function to be continuous at 500 m, for example,

$$B_1(\text{surface}) = B_0(\text{surface}) + g\int_0^{500} z\partial(\sigma_\theta - \sigma_1)/\partial z dz \qquad (2)$$

Therefore, the linear equations relating station surface Bernoulli functions for matchings from below 500 m will contain additional known terms, as a consequence of (2).

Killworth and Bigg (1988), in an intercomparison of inverse methods using time-averaged data from an eddy-resolving general circulation model, found that in most regions of the ocean the Bernoulli method performed, in a statistical sense, better than the other inverse methods tested in predicting the velocity field. This was despite the fact that diffusive processes are not included in the Bernoulli method formulation, while in the beta spiral, for instance, these may be considered (if somewhat unsatisfactorily, see Bigg, 1985). They also showed that a useful indicator of the accuracy of an inversion using the Bernoulli method was that the normalized rms deviation be less than about 0.08. We shall use this criteria in the assessment of the inversions. Note that McDougall (1989) feels there is an intrinsic error of nearly 0.5 cm s^{-1} in choosing potential density rather than neutral surfaces in the method. In allowing the density variable to alter character with depth, and in using the Bernoulli method over limited depth ranges, it is hoped to lessen this problem. In the data under study the flows are, in any case, substantially above this speed.

4. Bernoulli inversions

A number of trials using different formulations of the Bernoulli method were carried out, where the density variable used was altered, or limits placed on the depth range over which the searching for crossing points was performed. In addition, a trial where the full depth range was used, but the reference potential density was allowed to vary discretely (as discussed in Section 3), was carried out. Table 1 gives the normalized rms deviation for these trials. It shows that the most reliable are those using deeper data only, probably because of the frontal nature of the upper level flow, which is both ageostrophic and possibly less temporally stable than deeper flows.

The velocity fields at selected depths for the region 24.5–22.5W, 32.5–34.5N are shown in Figure 6, from the trial using σ_2 over 1000–3000 m. Note that this depth range means that velocities above 1000 m are effectively only due to thermal wind shear as the information used by the Bernoulli inversion ignores the probably nonadiabatic upper layers. Despite the scatter in the rms deviations in Table 1 all inversions, because of the vertical shear, give similar velocity fields for the surface layers, the major differences being below 700 m. The inversions actually produce surface pressure fields at the stations shown in Figure 1. The pressures, evaluated from the hydrostatic equation for the selected depths, were then interpolated onto a regular 1° × 1° grid over

Table 1. Normalized rms deviations for Cruise 138 inversions. σ_n denotes an inversion with variable σ (see text).

Density variable (depth range – m)	normalized rms deviation
σ_n (full)	0.129
σ_θ (full)	0.136
σ_θ (0-1000)	0.153
σ_1 (0-2000)	0.150
σ_2 (1000-3000)	0.025
σ_3 (2000-4000)	0.008
σ_4 (3000-5000)	0.005
σ_n (full)*	0.157
σ_θ (full)*	0.138
σ_1 (full)*	0.153

*These inversions used a potential vorticity calculated from the buoyancy frequency rather than $f\partial\sigma/\partial z$.

25–22W, 32–35N. The velocities shown in Figure 6 were derived from a finite differentiation of the geostrophic equations at the center points of the pressure grid.

Figure 5 shows a plot of σ_θ versus potential vorticity for the ten stations to illustrate the crossing point problem. The most useful information (that is, linear equations which are essentially independent) tends to come from the middle part of the water column. In the deeper water the curves tend to be similar and crossing points are likely to occur due to noise in the data rather than because of a common streamline passing between the points in question. It is therefore important to use a dynamically informative density variable in the deeper waters in order to reduce noise contamination.

There are several prominent features in Figure 6. At 100 db there is a strong eastward jet reaching speeds of more than 15 cm s^{-1}, which extends down to 500 m, in the northern part of the study area. In the rest of the area the general upper level climatological flow (see Fig. 1) is distorted toward the south, suggesting a frontal region to the east of the area (Fig. 6a). At 1000 db (Fig. 6b) the flow in the south is to the east, as opposed to the northwest in the climatology. An anticyclonic circulation around the region of less dense water shown in Figure 4b is found in the northwest of the region. The flow at 1500 db and below is weak, but at 1500 db (Fig. 6c) follows the σ_1 contours southward in the west of the region. Little evidence of a counter-eddy in the water affected by double diffusion below the MW is present, unlike the Tourbillon experiment (Arhan and de Verdiere, 1985).

The general picture given by the data and inversions is as follows. An upper level frontal zone around 34N separates a region of strong (>15 cm s^{-1}) eastward flow from the main, roughly climatological, flow in the south (see Figs. 2a, 3a, 4a, 6a). There appears to be an extension of this front to the south just to the east of our study area causing a veering of the climatological flow. This zone persists to a depth of 500 db,

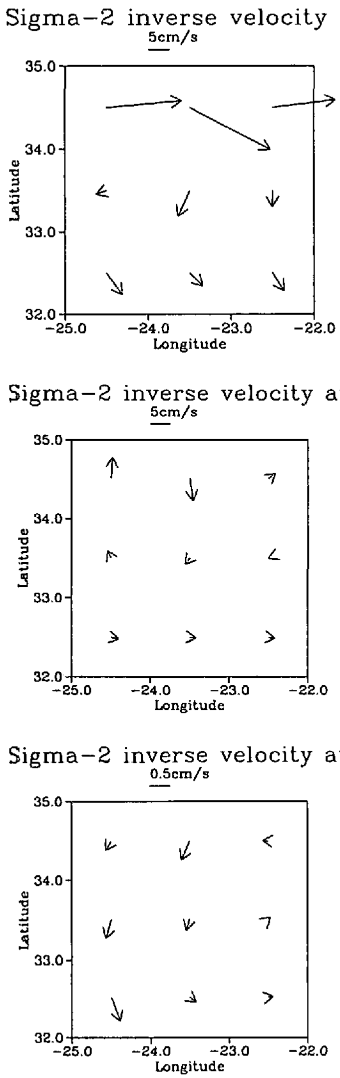


Figure 6. Velocities from Bernoulli method inversions using sigma-2 at three levels.

below which an intrusion of MW from the north becomes the dominant feature. The effect of this is to produce an anticyclonic eddy (Figs. 4b, 6b) near 1000 db, at the tip of the MW intrusion. Below the intrusion double diffusive processes have created a pool of warm salty water which acts to reverse the climatological northward flow of the area. In Figure 7 velocity hodographs with depth for two locations on either side of the frontal zone are shown. Here both the effect, and vertical extent, of the frontal zone are illustrated, as well as the anticyclonicity of the MW intrusion. Only near 1500 db do the two pictures give a similar velocity. Above, the horizontal velocity shear associated

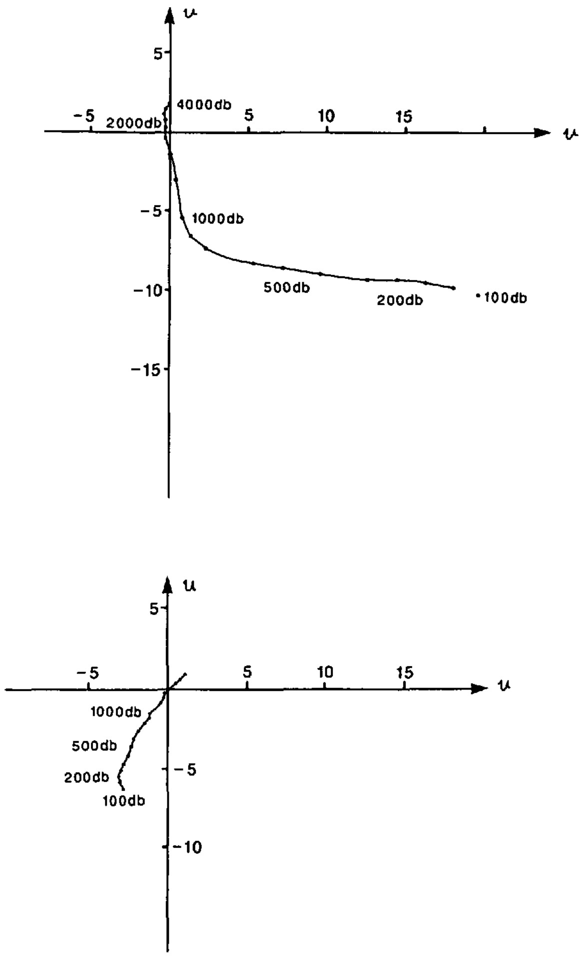


Figure 7. Velocity hodographs at two points on either side of the front. (a) 34.5N, 23.5W; (b) 33.5N, 23.5W.

with the front is strong. Below, the backing of the current from east to west shows the light density anomaly associated with the MW intrusion.

5. Comparison with other observations

Previous observations in this general area have demonstrated the occasional presence of a frontal zone, called the Azores Front. In addition, current meter data from the NEADS 1 site to the east of the study area are available for the period of the *Discovery* cruise. It is therefore possible to examine the consistency of the present conclusions with other knowledge.

From the dynamic topography for the region during early April 1982 Kase *et al.* (1985, Fig 9) found a strong zonal front meandering southward near 23W. This is the

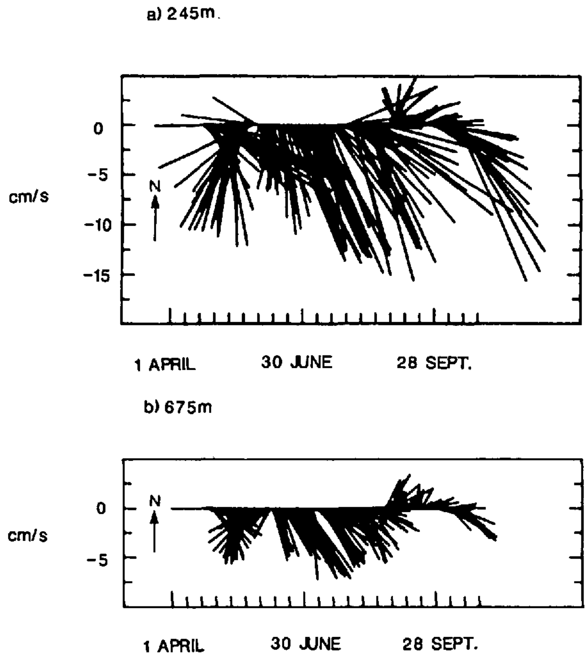


Figure 8. NEADS-1 current meter records for the mid-1983 deployment in the upper ocean at 2 depths (after Muller, 1984).

same region as the postulated frontal zone in the summer of 1983. Kase *et al.*'s results suggest that the 1982 front was a southward-thrusting loop in a bigger system. The data from the later *Discovery* cruise implies that the front was then farther to the east and extended farther southward. The persistence of the southward flow shown by the NEADS 1 current meter mooring (see Fig. 8) suggests that this southward extension was of several months duration. Current meter observations from the NEADS 1 area, and to the south (Muller and Zenk, 1983) support the dynamic topography's interpretation of the April 1982 manifestation of the front not penetrating farther south than shown in Kase *et al.* (1985).

Thus upper level frontal transitions of the scale observed in June 1983 are not unprecedented for this region. Indeed, the long-term current meter records at the NEADS 1 site at 22W, 33N (Muller, 1981; 1984) suggest that the front exists with this meridional orientation frequently, with a lifetime of several months. As the NEADS 1 site was occupied during the period of the *Discovery* cruise it is a valuable check on the velocity inversions. The acutal mooring site lies 50 km to the east of the region shown in Figure 6 so that the velocities are not directly comparable. Figure 8 shows strong southward currents at the upper meters of about 15 cm s^{-1} during June-July 1983, with the flow decreasing in strength dramatically with depth. This is consistent with our interpretation of the southward extension of the front being near the eastern extent of the study area. The deeper meters (1575 m, 2980 m—not shown) show a flow still to the south, although very weak (<1 cm s^{-1}). This is consistent with the inversion using σ_4 over 3000–5000 m. The discrepancy between σ_2 and σ_4 inversions is 0(0.5 cm s^{-1}), consistent with McDougall (1989).

6. Conclusions

General consistency between the CTD data of *Discovery* cruise 138, previous observations in the Madeira abyssal Plain, NEADS 1 current meter measurements and velocity inversions produced by the Bernoulli method has been demonstrated. An upper level frontal zone with a southward excursion was present during the summer of 1983, which the NEADS 1 data suggests persisted over several months. A current shear across the front at 100 db of some 20 cms^{-1} is suggested by the inversions. This shear lessens with depth to disappear near 700 db. At this level a loop of MW from the north leads to the development of an anticyclonic circulation near the tip of this intrusion. Penetration of warm salty water to greater depths results in a reversal of the climatological flow beneath the intrusion itself.

The thermal wind shear in the data is so large that the upper front would be obvious from geostrophy alone. However, the Bernoulli inverse method is able to provide, in addition, useful information on the deeper flow pattern associated with the MW intrusion.

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REFERENCES

Arhan, M. and A. C. de Verdiere. 1985. Dynamics of eddy motions in the eastern North Atlantic. J. Phys. Oceanogr., *15,* 153–170.

Bigg, G. R. 1985. The beta spiral method. Deep-Sea Res., *32,* 465–484.

Bigg, G. R. and P. D. Killworth. 1988. Conservative tracers and the ocean circulation. Phil. Trans. Roy. Soc. A., *325,* 177–187.

Gould, W. J. 1983. RRS *Discovery* 138, 17 June–9 July 1983. IOS Cruise Rep. No. 152, 25 pp.

Harvey, J. and M. Arhan. 1988. The water masses of the central North Atlantic in 1983–84. J. Phys. Oceanogr., *18,* 1855–1875.

Kase, R. H. and G. Siedler. 1983. Meandering of the subtropical front southeast of the Azores. Nature, *300,* 245–246.

Kase, R. H., W. Zenk, T. B. Sandford and W. Hiller. 1985. Currents, fronts and eddy fluxes in the Canary Basin. Prog. Oceanogr., *14,* 231–257.

Killworth, P. D. 1986. A Bernoulli inverse method for determining the ocean circulation. J. Phys. Oceanogr., *16,* 2031–2051.

Killworth, P. D., and G. R. Bigg. 1988. An intercomparison of inverse methods using an eddy-resolving general circulation model. J. Phys. Oceanogr., *18,* 987–1008.

Levitus, S. 1982. Climatological Atlas of the World Ocean. NOAA Tech. Paper 3, 173 pp.

Long, B. 1985. Determining the Bernoulli function from hydrographic data. Ocean Modelling, *63,* 8–10. (unpublished manuscript).

Maillard, C. 1986. Atlas hydrologique de l'Atlantique Nord-Est, IFREMER, Brest, 165 pp.
McDougall, T. J. 1987. Neutral surfaces. J. Phys. Oceanogr., *17*, 1950–1964.
—— 1989. Streamfunctions for the lateral velocity vector in a compressible ocean. J. Mar. Res., *47*, 267–284.
Muller, T. J. 1981. Current and temperature measurements in the North-East Atlantic during NEADS. Berichte Inst. Meer. Christian-Albrechts-Universitat Kiel, *90*, 100 pp.
—— 1984. Eulerian current measurements from the North East Atlantic March 1982–October 1983. Berichte Inst. Meer. Christian-Albrechts-Universitat Kiel, *127*, 99 pp.
Muller, T. J. and W. Zenk. 1983. Some Eulerian current measurements and XBT-sections from the North East Atlantic October 1980–March 1982. Berichte Inst. Meer. Christian-Albrechts-Universitat Kiel, *114*, 145 pp.
Pollard, R. T. and S. Pu. 1985. Structure and circulation of the upper Atlantic Ocean northeast of the Azores. Prog. Oceanogr., *14*, 443–462.
Robinson, M. K., R. A. Bauer and E. H. Schroeder. 1979. Atlas of North Atlantic-Indian Ocean monthly mean temperatures and mean salinities of the surface layer. NOO, Rep-18, 234 pp.
Saunders, P. M. 1987. Flow through *Discovery* Gap. J. Phys. Oceanogr., *17*, 631–643.
Saunders, P. M. and F. Dolan. 1987. CTD Data from *Discovery* Gap and the Madeira Abyssal Plain. IOS Rep., No. *236*, 76 pp.
Welander, P. 1971. The thermocline problem. Phil Trans. Roy. Soc. London A, *270*, 415–421.
—— 1983. Determination of the pressure along a closed hydrographic section: 1. the ideal case. J. Phys. Oceanogr., *13*, 797–803.
Wust, G. 1935. Scichtung und Zirkulation des Atlantischen Ozeans. Die Stratosphare, *in* Wissenschaftliche Ergebnisse der Deutshen Atlantischen Expedition auf dem Forschungs und Vermessungsschiff "Meteor" 1925–1927, 6. Lieferung 1. Teil, 2, 180 pp.

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