International Council for the Exploration of the Sea

C.M.1977/C:32
Hydrography Committee

EVIDENCE FOR A POLEWARD EASTERN BOUNDARY CURRENT IN THE NORTH ATLANTIC OCEAN

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INTRODUCTION

Currents in the deep water near the continental slope of the Eastern North Atlantic tend to have a poleward component, at least between latitudes 37°N and 53°N. Mean speeds are typically a few centimetres per second. The region of poleward flow is not confined to the core of the Mediterranean water, but is much thicker and may indeed extend through most of the water column. Its width is not well defined by the existing observations, but off the Iberian Peninsula it may be appreciably wider than the region of relative northward geostrophic flow in the Mediterranean core immediately adjacent to the slope.

Evidence for such a boundary current has been accumulating for several years, but its existence as a permanent continuous feature has seemed doubtful (and still does) in view of the variable nature of deep currents measured farther offshore. In mid-ocean, typical speeds in the deep water averaged over several days are also a few centimetres per second, but they vary with spatial scales of several tens of kilometres and time scales of several weeks. In such a background, weak features of the larger scale longer term circulation are hard to detect, when only sparse measurements of all too short duration are available. We became convinced of the probable reality of this boundary current by some float trajectories observed in January-February 1977. They are described below first, and then other more extensive evidence is reviewed.

DIRECT MEASUREMENTS OF DEEP CURRENTS ALONG A ZONAL SECTION BETWEEN THE MID-ATLANTIC RIDGE AND SPAIN

In January-February 1977 the RRS 'Discovery' occupied a section passing through two of the NEADS* mooring sites, from the Mid-Atlantic Ridge to the continental slope off Spain, between latitudes 41°N and 42°12'N (Fig. 1). Forty CTD stations were occupied, and neutrally buoyant floats were tracked at nineteen locations. The section was worked in three parts. The five westernmost floats were tracked first, and the related CTD stations occupied. The ship then set off for Vigo, launching the six easternmost floats, and working CTD stations near them, on the way in. Those floats were located and recovered on leaving Vigo, and intermediately spaced CTD stations were occupied. Then the middle of the section was filled in with eight more float tracks and seventeen CTD stations. All the floats were loaded for 1500 m depth. For most of the tracks, only starting positions (after sinking) and end points are available. Provisional details are given in Table 1. Positions depend on interpolation between satellite fixes using a 2-component e.m. log and gyro. For a track lasting 5 days (the average duration) the error in mean velocity due to errors in navigation and to truncation of inertial and semidiurnal tidal oscillations is unlikely to exceed 0.2cm/sec. For the shortest trajectory, relative fixing was much more accurate because bottom transponders were used, and the effect of semidiurnal tide (the predominant oscillation) was minimized by tracking for 25 hours.

*NEADS = North-East Atlantic Dynamics Studies

How representative in time are these relatively short samples of current? For a two-month current meter record at 1500 m from the NEADS site near 42°N, 14°W, the r.m.s. difference between 30-day means and the 5-day means contained within them was 0.6 cm/sec east-west and 0.7 cm/sec north-south; the magnitude of the two-month mean velocity was 1.6 cm/sec. This suggests that, to within about 0.7 cm/sec we may regard the components of float velocities as synoptic and representative of the four-week period during which the section was occupied.

Since the float trajectories were spaced about 90 km apart along the section, we can regard their north-south components as spatially independent samples in the variable mid-ocean current field. In MODE*, the transverse velocity correlation function had its first zero at 50 km, and from the little evidence that we have there is no reason to think that the scale is any larger in the Eastern North Atlantic.

The mean of the 19 samples of the component of current through the section is northward, 0.6 (±2.5) cm/sec (s.d. of a single observation) or ±0.6 cm/sec (s.d. of the mean). For the three easternmost samples, the mean is 2.7 (±1.4) cm/sec, still using 2.5 cm/sec as the standard deviation of a single observation, and the mean of the four westernmost values is -1.6 (±1.3) cm/sec. On the face of it, there was significant northward flow at 1500 m at the eastern end of the section and a less definite indication of southward flow at the western end. At best, however, this is only a synoptic view, valid for at most a few weeks in January-February 1977.

Except at the two mooring positions, currents were not measured directly at any other levels along the section. At

*MODE = Mid-Ocean Dynamics Experiment

the eastern end, the vertical structure of the possible boundary current has to be inferred by fitting the geostrophic profiles of relative current calculated from pairs of CTD stations to suitable mean values of observed current at 1500 m. How should these mean values be calculated? Since the observations are about 90 km apart, and the transverse correlation function is assumed to drop to zero at 50 km, almost nothing is known about the current midway between the pairs of float tracks. A rough but plausible way of obtaining mean values along the section would be to assign suitable estimates at the mid points and interpolate linearly between the successive observed and estimated values. For the estimates, one could use 0.6 ($\pm$2.5) cm/sec, the mean value for the whole section with the s.d. of a single observation, if one regarded all the observations as random samples of a variable velocity field. Or one could fit a linear east-west gradient as well as a mean value, and interpolate values from that, but a linear gradient does not seem oceanographically likely. Or, if one believed strongly in an eastern boundary current, one could use the mean of the three easternmost observations as an intermediate estimate. The first of these alternatives is the least biased in favour of a boundary current and has been used in choosing reference values for the geostrophic profiles in Fig. 2. Allowing for an error of approx. $\pm$1.2 cm/sec on the reference velocities used, the weak southward flow at 600-1200 m in the middle profile seems insignificant, and the net northward velocity does not vary much with depth. In particular the only sign of a northward maximum associated with the core of the Mediterranean water at about 1000 m is the very weak indication in the easternmost profile. The net northward transport derived from these three profiles is 11 ($\pm$ approx. 6) million m^3/sec. The width of the zone of northward flow, measured from the 1000 m depth contour out to station 9386, is 245 km.

OTHER CURRENT MEASUREMENTS NEAR THE EASTERN BOUNDARY

Other observations (all that we have been able to find) at 1500 m and a little deeper near the eastern boundary of the North Atlantic are shown in Fig. 3, added to those already described. Together they make a fairly convincing picture of a poleward boundary current. From south to north the added observations are: (1) a 5-day mean current of 6 cm/sec towards 351° at 2046 m, in a water depth of 2450 m (Meincke, Siedler and Zenk 1975): (2) the vector mean of 19 days of float tracks at 2560 m, 1.0 cm/sec towards 354° (Gould 1971): (3) the mean of 170 days of records from near-bottom current meters in approximately 2000 m water depth (data from NIO-IOS moorings, see Table 2): (4) a 29-day mean current of 5.7 cm/sec towards 036° at 1594 m, in 3118 m of water (personal communication from Dr I.D. James, IOS Bidston). One may well be doubtful whether even a 30-day mean has any longer-term significance, but the steadiness of the mean values of the individual records making up that 170-day series at position (3) is encouraging (see Table 2).

Not only at 1500 m and below, but at other levels at these sites, the observed mean currents tend to be poleward. Meincke, Siedler and Zenk (1975) had mean currents slightly east of north at four other levels, increasing downwards from 10 cm/sec at 234 m to 12 cm/sec at 744 m. A drogue at 1000 m, tracked for 2 days near 42°34'N, 10°24'W, between Galicia Bank and Spain, moved towards 003° at 5.7 cm/sec in winds that were mainly from the northwest (Lacombe 1961). That was interpreted as northward flow in the core of the Mediterranean water, but could equally well result from similar movement in a much thicker layer. In the mid-Biscay measurements discussed by Gould (1971), 48-day mean currents at two shallower levels were generally northward, 6.6 cm/sec, 024° at 350 m and 2.0 cm/sec, 045° at 1350 m. In the north Biscay area marked (3) in Fig. 3, three deep water sites have been occupied, two of them more than once. Mean values at each site, grouped by depths, are listed in Table 2 and illustrated in Fig. 4. At site A, the one with most observations, the mean vectors at all levels are to the northwest and remarkably uniform in magnitude, considering the great differences in averaging times at different levels. The shallowest series (310 m) is the least stable, as might be expected. The three samples at 972-1001 m are remarkably steady. The mean values at the other two sites are also consistently northward and mainly northwestward, though based on shorter records. At the 29-day Bidston mooring referred to above (position 4 in Fig. 2), generally northward mean currents were found at the other two levels from which results are available, 5.5 cm/sec, 043° at 1087 m and 5.2 cm/sec, 346° at 2550 m.

We can find very little additional evidence about the width of this boundary current. A group of three float tracks at 1800 m depth (Swallow 1972) offshore from the north Biscay mooring site suggest that at that time its width there could not have exceeded 60 km outwards from the 1000 m depth contour. But they provide only a 3-day sample. With that width, and taking a mean velocity of 3 cm/sec in the boundary current at all depths below 300 m, its transport there would have been only 3 million m^3/sec.

DISCUSSION

Although the current arrows in Fig. 2 make a consistent picture of a poleward boundary current, there are not many of them, and some readers may well be sceptical about their significance. Can we expect currents averaged over such short periods to represent a steady state? In some parts of the ocean we certainly cannot. For example, in a four-month record at 1995 m from NIO mooring 117 (52°29'N, 17°45'W) the mean current for the first half was 7.8 cm/sec, 008°, and for the second half 6.4 cm/sec, 228°. However, it seems possible that shorter averaging times are sufficient near the continental slopes. In a study of mean currents at WHOI site D, in 2600 m depth on the continental slope south of Cape Cod, Webster (1969) analyzed 25 records from various depths and varying in length from 19 to 60 days. All but one had a westward component of velocity. The same consistent behaviour can be seen in the mean values at the north Biscay site, listed in Table 2. In the records from an 8-month array of moorings on the continental rise at 69°-70°W, Luyten (1977) found steady westward flow above the upper rise, in water depths less than 3000 m, becoming progressively more variable towards deeper water. Looking at the observations plotted in Fig. 2, it is quite possible that the northward current over the deep abyssal plain in the centre of the Bay of Biscay was coincidental, but the others, near the continental slope, seem likely to be at least in the same general direction as the mean.

With so few observations, continuity of a boundary current must be uncertain. To the west of the Iberian peninsula, the northward spread of the core of Mediterranean water at 800-1200 m has been known for a long time (see, for example, Plate 56 in Helland-Hansen and Nansen, 1926), and is consistent with a continuous northward flow there. Hitherto, this seems to have been regarded as confined to the Mediterranean water only; Madelain (1967) found weak northward geostrophic flow in the Mediterranean water relative to 2000 dbar, close to the Iberian continental slope; Meincke, Siedler and Zenk (1975) interpreted their results as indicating weak relative northward advection in the upper Mediterranean water and presumably regarded the rest of their observed mean currents as transient. In our view, the core of northward-moving Mediterranean water close to the slope is embedded in a thicker and wider body of water also moving northward.

Beyond Cape Finisterre, Madelain (1972) found some evidence for eastward relative geostrophic flow into the Bay of Biscay, associated with the Mediterranean salinity maximum close to the continental slope off northern Spain at 8°W, but there were more pronounced maxima and stronger relative flow elsewhere offshore. In the Bay of Biscay itself, Le Floch (1969) deduced a generally cyclonic pattern of relative motion in the Mediterranean water, except for a small anti-cyclonic eddy in the southeast corner. At depths between 2000 and 2500 m, Worthington and Wright's (1970) charts of salinity distribution at potential temperatures between 4° and 3°C are consistent with a poleward eastern boundary current. But there is no firm evidence, and at present it is not clear how, or whether, the northward boundary current off the Iberian peninsula is connected to the observed current along the slope in the northern part of the Bay of Biscay.

It would be easier to believe in a continuous boundary current if there were some reason to expect it to be there, if it fitted some model. One might think of accounting for the northward flow off Portugal in the same way as the poleward undercurrent in the upwelling region off northwest Africa, as observed by Mittelstaedt, Pillsbury and Smith (1975) at 21°40'N. There is a mean southward wind stress off Portugal, and strong evidence of upwelling in the summer months (Wooster, Bakun and McLain, 1976). But in the northern part of the Bay of Biscay the mean wind stress is directed slightly north of east. According to Bunker and Worthington's (1976) charts of annual mean wind stress, the change of sign of the northward component near the eastern boundary occurs near Lat. 47°N. Some other explanation seems to be needed for the poleward current north of that latitude.

SUMMARY

Direct measurements of subsurface currents near the continental slope of the eastern north Atlantic are consistently poleward, with mean velocities of a few centimetres per second. The poleward flow has been found at all depths sampled, from 300 m to 3000 m. Two estimates of the width of the poleward current are 60 km and 245 km. Its continuity is not well established. At present, it seems best to regard the idea of a poleward eastern boundary current as a working hypothesis, to be substantiated or disproved by more observations.

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TABLE 1

Provisional data for float trajectories at 1500 m, Jan-Feb 1977

Date & Time of first fix		Start Pos. Lat. N.	Start Pos. Long. W.	Duration (hrs)	Mean Velocity cm/sec	Mean Velocity °T
17.1	2108	41 33.8	28 45.4	81.9	3.2	139
18.1	0530	41 25.5	27 57.3	79.9	0.9	221
18.1	1320	41 17.6	27 03.3	90.8	2.3	242
18.1	2154	41 07.7	26 05.0	89.5	2.4	162
23.1	0736	41 01.9	24 55.7	25.5	1.9	048
7.2	2100	41 04.3	23 52.4	30.2	2.4	111
7.2	1200	41 10.9	22 53.9	48.9	6.4	284
7.2	0206	41 16.8	21 48.2	83.9	3.5	180
6.2	1506	41 24.4	20 40.5	118.0	6.5	327
6.2	0254	41 30.3	19 35.8	147.9	4.2	328
5.2	1354	41 35.9	18 31.3	170.7	1.6	123
5.2	0242	41 42.2	17 27.2	197.5	1.0	040
4.2	1718	41 47.9	16 22.3	215.6	0.9	078
26.1	1054	41 53.0	15 17.9	196.1	1.5	020
26.1	1916	41 55.3	14 07.3	180.0	2.3	337
27.1	0254	41 59.9	13 02.9	156.2	1.8	176
27.1	1218	42 05.7	11 53.2	137.2	4.0	337
27.1	2112	42 06.6	10 41.1	103.6	1.8	336
28.1	0256	42 09.5	9 51.1	84.0	2.8	340

TABLE 2

Mean currents from moorings at sites in the northern part of the Bay of Biscay

(A) Near 47°35'N, 8°25'W (2100-2200 m depth)

Start Date	Depth (m)	Duration (days)	E.compt. (cm/sec)	N.compt. (cm/sec)	Total Days	Depth	Weighted mean E.compt.	Weighted mean N.compt.
6. 9.68	368	44	-10.0	8.3				
25. 3.69	257	70	- 0.76	0.51				
4. 6.70	304	21	12.0	1.6				
5. 2.71	292	21	- 4.44	-0.60	330	310	-2.66	2.44
19. 6.71	338	51	- 3.72	-2.49				
7. 8.71	312	49	- 4.02	0.24				
26. 9.71	311	74	- 2.11	6.73				
9. 1.71	469	23	- 1.10	1.40				
5. 2.71	492	21	- 4.76	-0.15	44	480	-2.85	0.66
5. 2.71	990	21	- 2.49	1.33				
19. 6.71	1001	31	- 3.43	1.23	101	985	-3.06	1.28
7. 8.71	972	49	- 3.07	1.30				
5. 2.71	1487	21	- 1.52	1.69				
6. 5.72	1649	9	- 1.54	3.34	30	1536	-1.53	2.19
5. 2.71	1986	21	- 3.92	0.40				
4. 4.71	1988	24	- 2.17	-0.14				
19. 6.71	2021	51	- 1.04	0.27	170	1995	-1.76	0.15
26. 9.71	1981	74	- 1.52	0.08				

(B) Near 47°45'N, 8°05'W (1000 m depth)

Start Date	Depth (m)	Duration (days)	E.compt. (cm/sec)	N.compt. (cm/sec)	Total Days	Depth	Weighted mean E.compt.	Weighted mean N.compt.
6. 6.69	306	41	- 0.41	-1.32				
15. 6.71	226	49	2.15	0.33	122	270	1.54	0.22
13. 8.71	293	32	3.11	2.03				
13. 8.71	957	44	- 2.18	4.31				

(C) 47°18'N, 7°40'W (4156 m depth)

Start Date	Depth (m)	Duration (days)	E.compt. (cm/sec)	N.compt. (cm/sec)
30. 1.71	110	27	- 9.4	4.2
"	526	27	- 4.7	3.2
"	1100	27	- 1.2	3.6
"	2121	27	- 0.5	2.0
"	3141	27	0.0	1.4

APPENDIX: ESTIMATING THE REFERENCE VELOCITY FOR HYDROGRAPHIC STATIONS

Given observations of the flow (assumed averaged over tidal and inertial periods) at a number of locations but at the same depth how is the reference velocity for a pair of stations in the vicinity to be computed?

When the stations and current measurements are made in the same line only the flow component normal to that line need be considered. A simple linear interpolation formula is then

$$\bar{V}_{est} = \sum_{i=1}^{N} \alpha_i (V_i + \epsilon_i) \quad (1)$$

where $V_i + \epsilon_i$ is the observed $\perp$ velocity, a combination of true velocity V_i and random error ϵ_i, α_i is the weight (to be determined) and $\bar{V}_{est}$ is the estimate of the spatial average between the hydrographic stations (the reference velocity). The mean value ('local' or 'large scale') of the flow is supposed removed from the data so that a number of attempts to estimate $\bar{V}$ yields $<\bar{V}_{est}> = 0$. The mean squared error E^2 of the estimate determined from these attempts however is non zero,

$$E^2 = <(\bar{V} - \bar{V}_{est})^2> \quad (2)$$

where $\bar{V}$ is the true mean. Optimum linear interpolation (see Bretherton, Davis and Fandry, 1976) minimises E^2 by appropriate choice of α_i, viz $\frac{\partial E^2}{\partial \alpha_i} = 0$, which yields by substituting (1) in (2)

$$0 = -<V_i\bar{V}> + \sum_{k=1}^{N} \alpha_k (<V_iV_k> + \delta_{ik}\epsilon^2) \quad (3)$$

a set of N simultaneous equations for α_i. Substitution of (3) in (2) give the minimum squared error of the estimate as

$$E^2_{min} = <\bar{V}^2> - \sum_{i=1}^{N} \alpha_i <V_i\bar{V}> \quad (4)$$

To compute α_i and E^2_{min} the following covariances must be known:- $<V_iV_k>$, $<V_i\bar{V}>$ and $<\bar{V}^2>$. Given a knowledge of the first the second and third follow. With Freeland and Gould (1976) we take the (transverse) velocity correlation function for two points separated by distance x as

$$g(x) = (1 + bx - b^2x^2)\exp(-bx) \tag{5}$$

where b is chosen to be $(31\ \text{km})^{-1}$, $g(x)$ then crossing zero at 50 km. The correlation between the $\perp$ flow at x = 0 and $\perp$ flow averaged between separations x_1 to x_2 may then be shown to be

$$g'(x_1,x_2) = \frac{b}{x_2-x_1}\left(x_2^2e^{-bx_2} - x_1^2e^{-bx_1}\right) + \frac{1}{x_2-x_1}\left(x_2e^{-bx_2} - x_1e^{-bx_1}\right) \tag{6}$$

The variance of the averaged flow $\overline{V}$ is determined by first computing the relative spectral density function of the fluctuating flow from (5);

$$F(k) = \frac{16k^2b^3}{\pi(b^2+k^2)^3}$$

where k is the wavenumber $(= {}^{2\pi}/x)$. Averaging flow over a distance $L = x_2 - x_1$ multiplies the spectral density function by $\frac{\sin^2(kL/2)}{(kL/2)^2}$ so the variance of the averaged flow compared with that of the local flow is

$$A(L) = \int_0^\infty F(k)\,\frac{\sin^2\left(\frac{kL}{2}\right)}{\left(\frac{kL}{2}\right)^2}\,dk \tag{7}$$

The expressions $\langle V_iV_k\rangle$, $\langle V_i\overline{V}\rangle$ and $\langle\overline{V}^2\rangle$ are obtained by multiplying equations (5), (6) and (7) by the variance of the fluctuating flow σ^2.

A sample calculation is shown in figure 5 of the weights α and error estimate E from the above analysis using four equally spaced current observations (with prescribed random error) to estimate the average flow between the central pair. Note that to this must be added the 'mean' originally subtracted from the data. It will be seen that 90 km spacing yields the most uncertain average flow estimate (not taking into account the uncertainty of the 'mean'): half the spacing with twice the observational error would yield a more certain estimate of the average needed to compute the reference velocity for a station pair.

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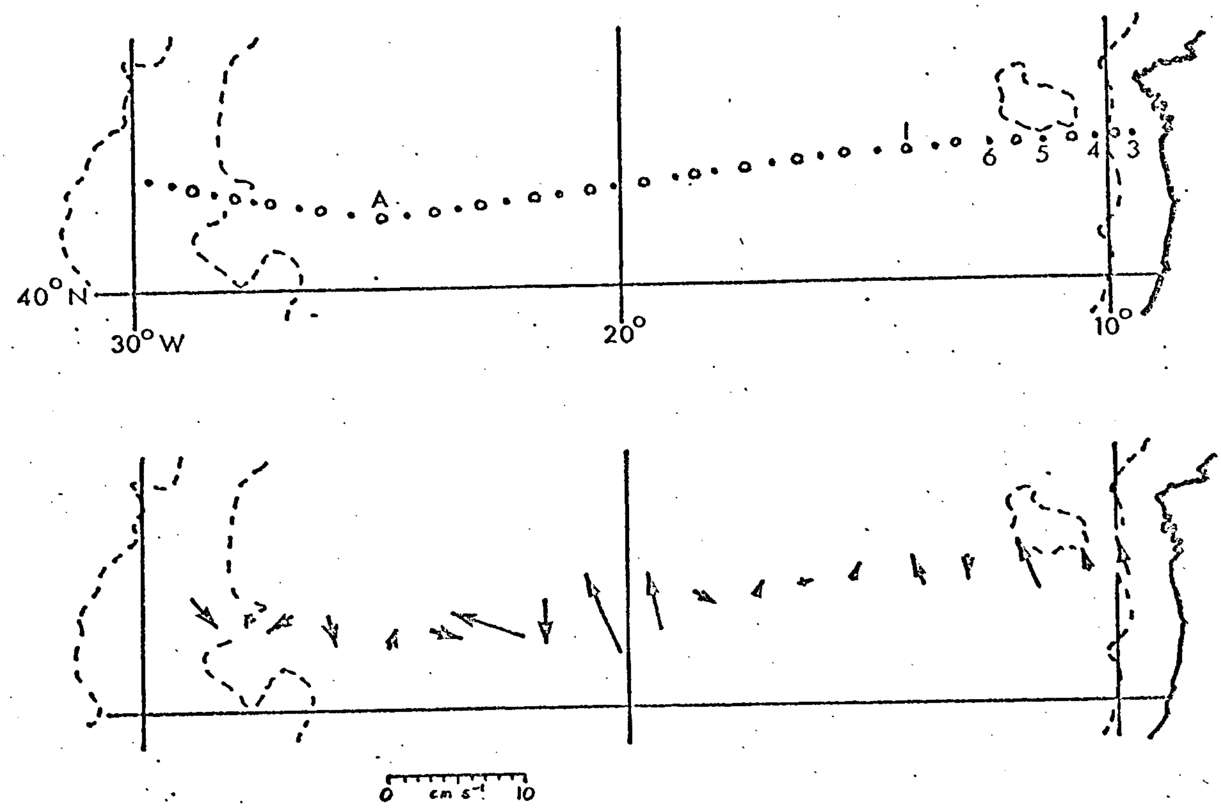

Figure 1. (upper) Section occupied by RRS 'Discovery' in Jan.-Feb. 1977. Solid circles are CTD station positions, open circles are CTD + float launch positions. A and I are NEADS mooring sites. 3,4,5,6 indicate positions of CTD stations 9383-6 used in geostrophic profiles in Fig. 3. The dashed lines are approximate 2000 m bathymetric contours.

(lower) Mean velocity vectors for 1500 m float trajectories.

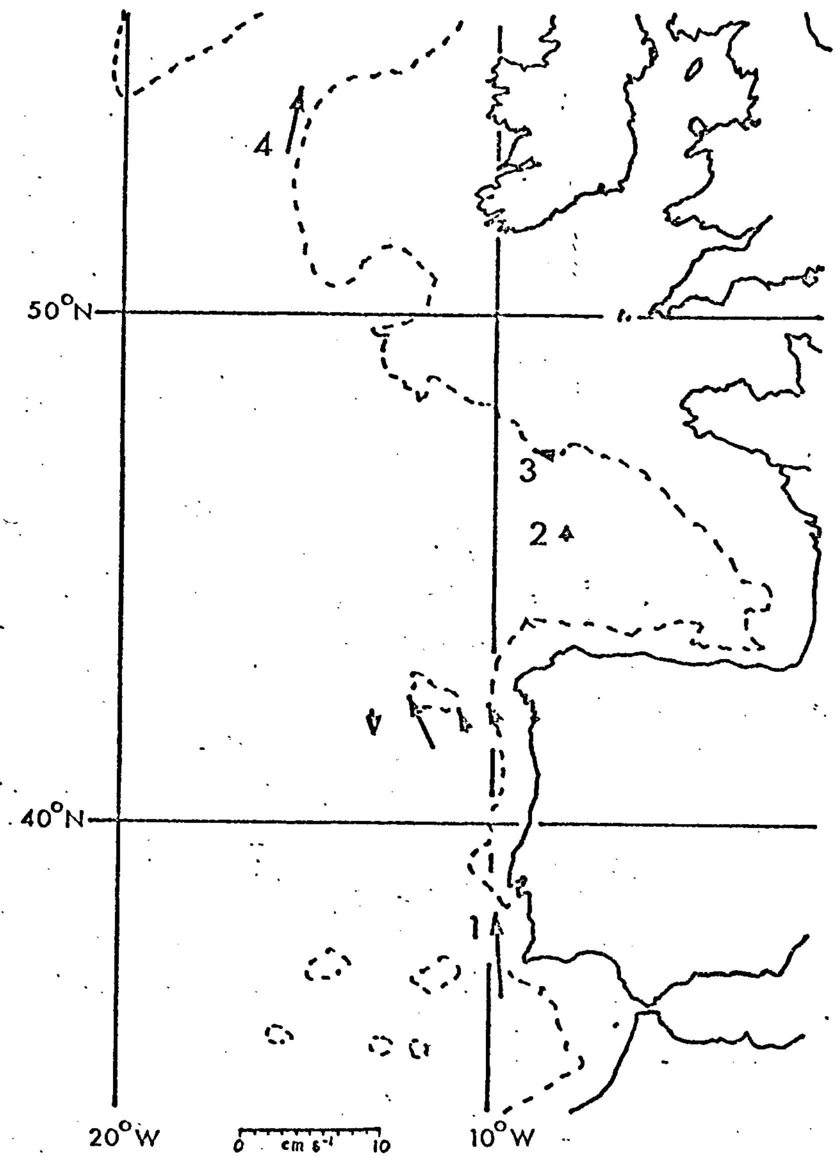

Figure 2. Mean velocities at 1500 m and deeper near the continental slope in part of the Eastern North Atlantic. Numbers are explained in text.

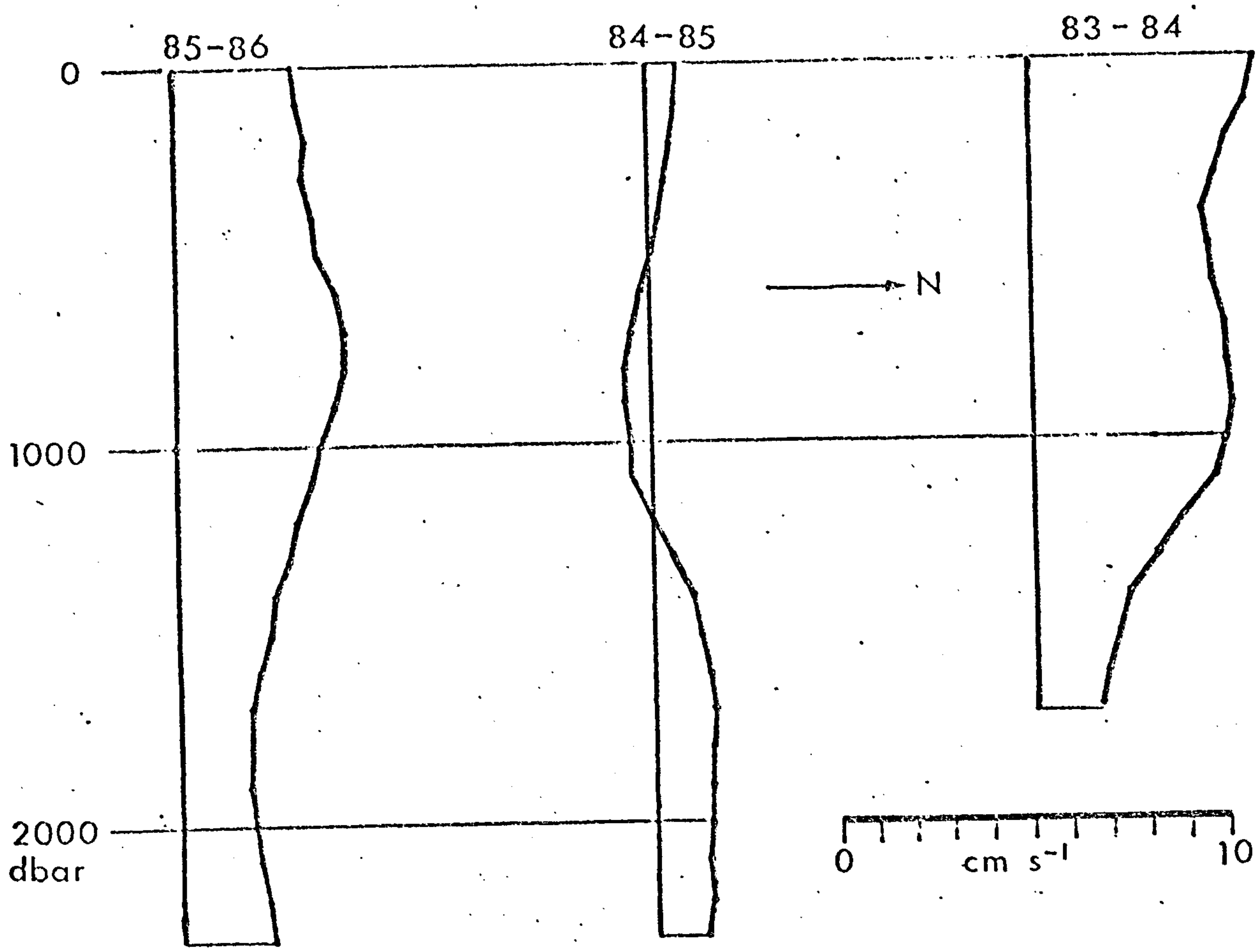

Figure 3. Vertical profiles of geostrophic current between CTD stations 9383-9386, fitted to observed mean current between each pair at 1500 m (1516 dbar).

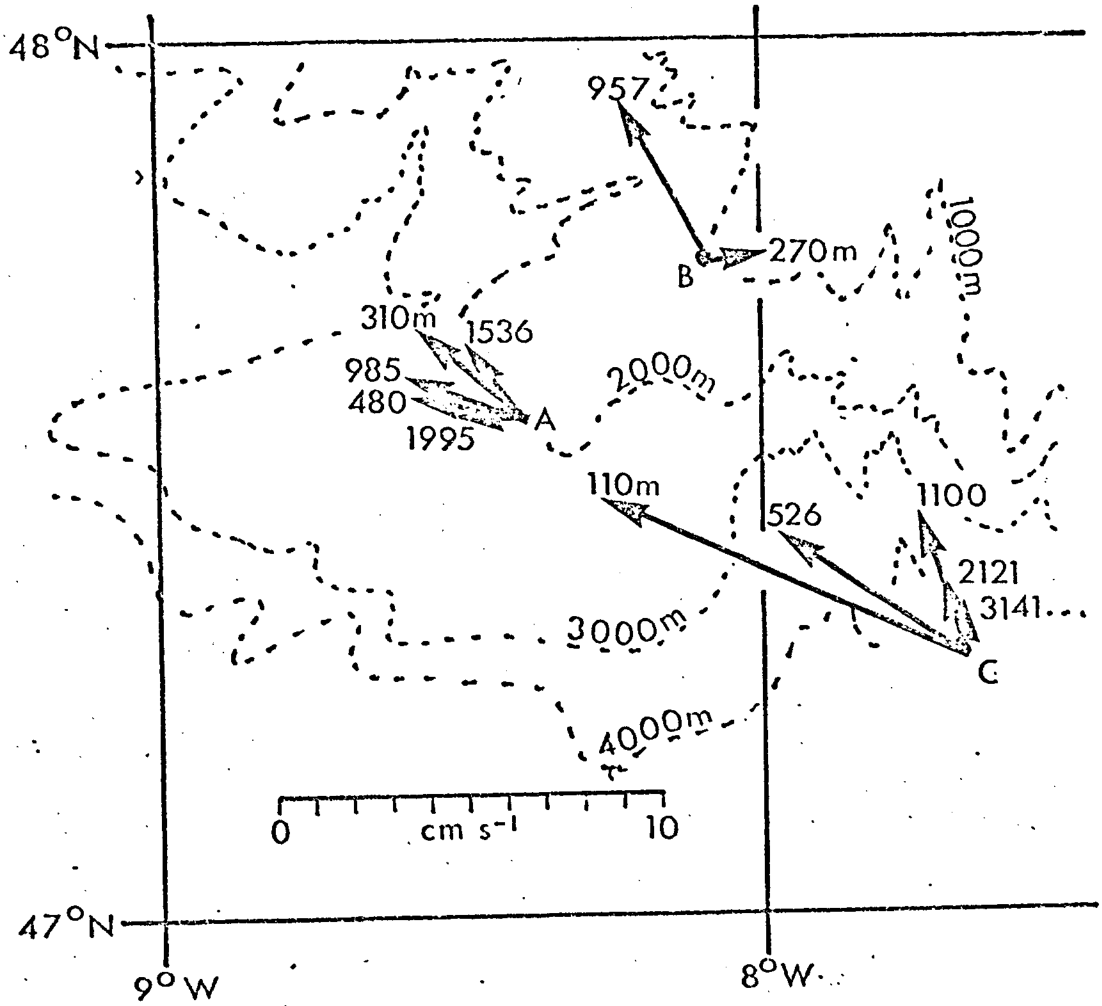

Figure 4. Mean velocities at three mooring sites in the northern part of the Bay of Biscay. These are the overall mean values for each site listed in Table 2.

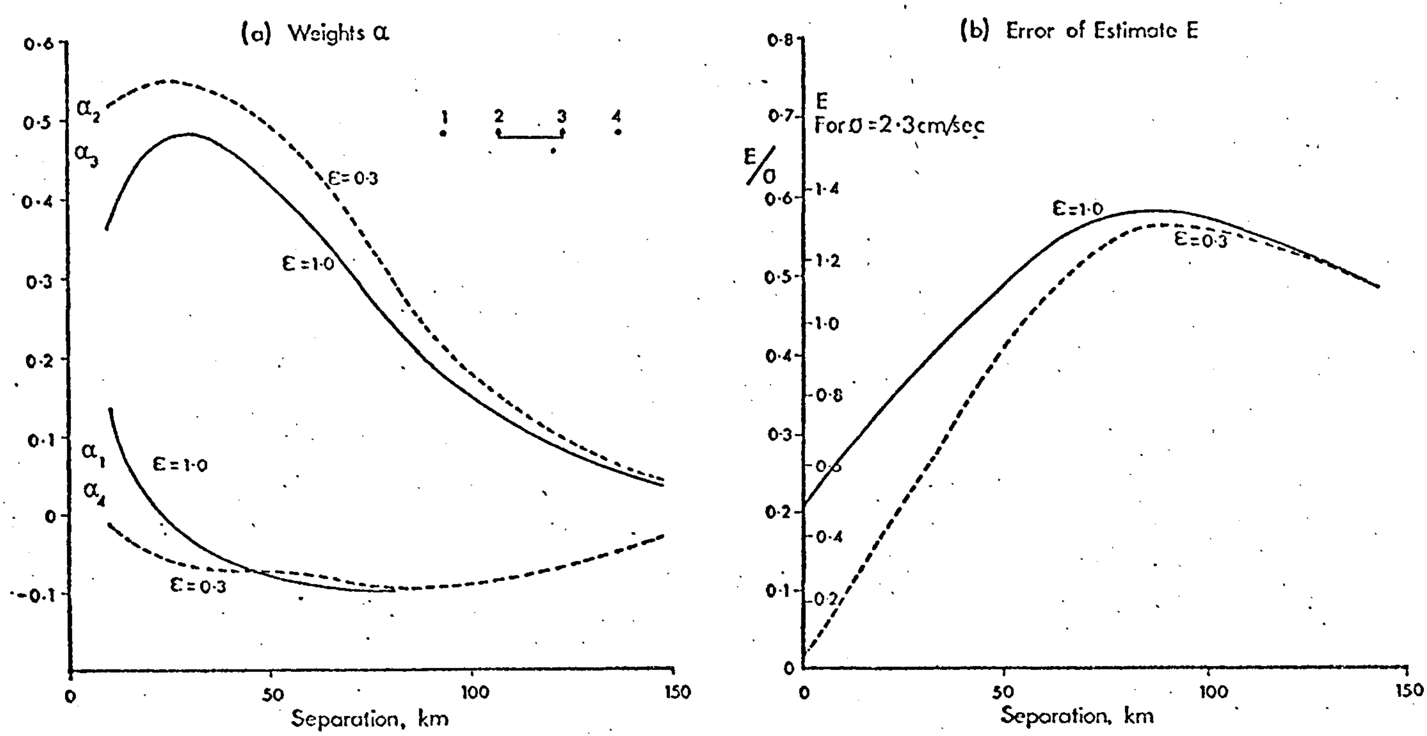

Figure 5. Estimation from four current observations of spatial average between central pair. ε is the error of an individual observation (cm/s) and σ is the r.m.s. fluctuation of the current. (Left) Weights and (Right) Error of spatial average estimate.