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Subsurface currents off Cape Hatteras*

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Abstract—In October 1962, direct current measurements at depths of 800 m to 2500 m beneath the zone of swiftest surface currents near Cape Hatteras revealed a subsurface flow to the south. Analyses of temperature-salinity and dissolved oxygen distributions indicate that the immediate source of the southward flowing water was the region east of Cape Hatteras between the Gulf Stream and the Continental Slope. These measurements seem to affirm that continuity exists between the deep westward flow observed in the Slope Water Region by VOLKMANN and the southward flow east of Cape Romain, near 33°N, reported by SWALLOW and WORTHINGTON. The geostrophic volume transport of the southward flow was between 4 million and 12 million m^3/sec. The surface of zero axial velocity was at a depth of about 500 m at the left side of the region of swift surface current (inshore), and deepened rapidly in the offshore direction.

INTRODUCTION

THE question of the nature of a deep southward flow along the west side of the Atlantic basin as described by WÜST (1936) has recently received considerable attention . STOMMEL (1957) revived interest in the problem by demonstrating that such a flow was logically predictable on theoretical grounds. The development of the neutrally buoyant float technique by SWALLOW (1955, 1957) provided a means for making sufficiently accurate deep current measurements to investigate this flow directly.

Using neutrally buoyant floats, SWALLOW and WORTHINGTON (1961) located a deep southward flow to the east of the Florida Current in latitude 33°N, and VOLKMANN (1962) on two occasions found a deep westward flow in the Slope Water Region (ISELIN, 1936) south of Cape Cod. However, during " Gulf Stream '60 " (FUGLISTER, 1963) direct current measurements under the principal slope of the main thermocline near the Kelvin seamount showed no reversal : deep floats moved eastward. These three locations are shown in the inset to Fig. 1.

These scattered current measurements, supporting in general the conclusions of Wüst and Stommel, suggested that there might be a direct link between the westward drift in the Slope Water Region and the deep southward flow east of Cape Romain. If this were the case, the deep countercurrent would be required to pass under the surface current, probably somewhere to the east of Cape Hatteras.

The influence of the bottom topography on the behaviour of the Florida Current immediately after it leaves the shallow (800 m) Blake Plateau is not known, but the general configuration of the sea bottom with respect to the mean surface current is noteworthy. As the Blake Plateau tapers off to the northeast, the Florida Current,

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which on the plateau extends to the ocean bottom (PRATT, 1963), flows along the steep gradients of the continental escarpment (HEEZEN, *et al.* 1959), and then at the latitude of Cape Hatteras away from the escarpment into the deep (5000 m) western basin. The effects on the current of the disappearence of first the shallow lower boundary and finally the continental escarpment, or lateral boundary, represent an interesting field of investigation, but present attention is focused on the idea that within this transition zone near Cape Hatteras the western boundary undercurrent may pass obliquely under the surface current.

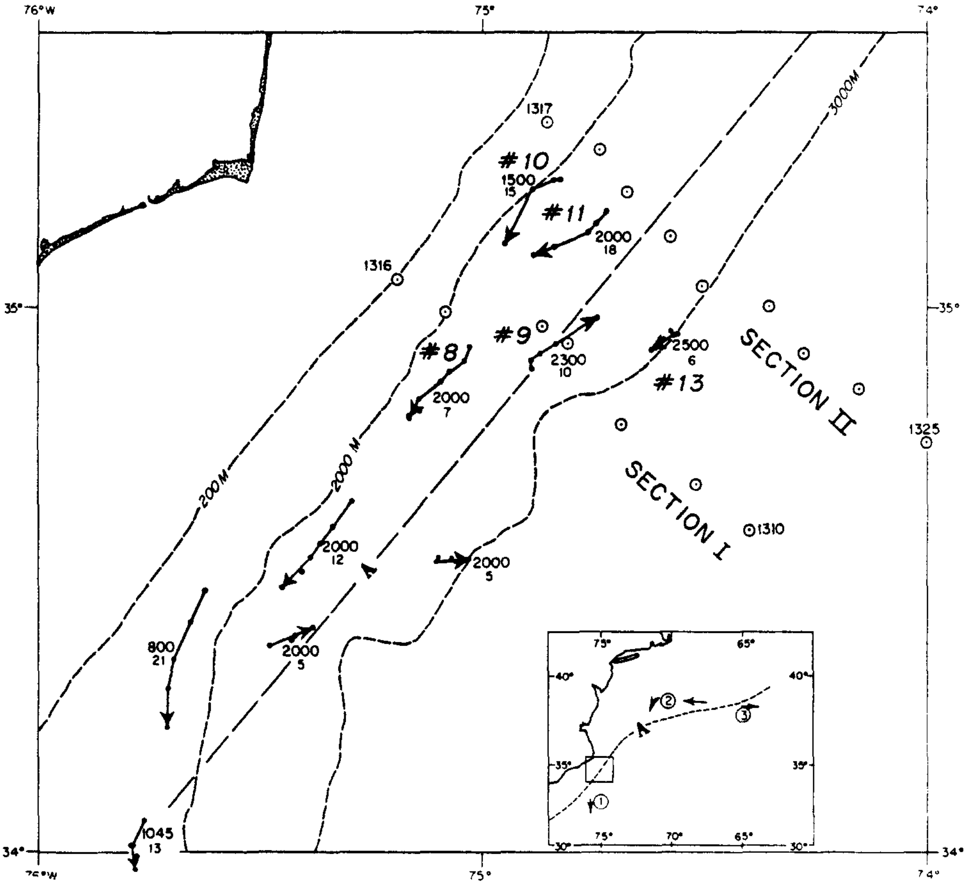

Fig. 1. Tracks of neutrally buoyant floats and locations of oceanographic stations. Numbers beside arrows indicate float depth (metres) and speed (cm/sec), and identify by number those floats which appear on the vertical sections. Line *A* represents the "approximate position of Axis of Gulf Stream" from H. O. 1411. Inset shows locations of some previous direct deep current observations by (1) SWALLOW and WORTHINGTON, (2) VOLKMANN and (3) FUGLISTER.

THE OPERATION

A combined operation by the research ships *Atlantis* and *Crawford* of the Woods Hole Oceanographic Institution was executed in the fall of 1962. *Atlantis* was engaged primarily in tracking neutrally buoyant floats. These were released at the positions shown in Fig. 1. In addition *Atlantis* attempted bottom current measurements using a BRUCE (1962) photographic current meter. On board *Crawford* the observations consisted of oceanographic sections (using both Nansen bottles and an *in-situ* salinometer) and numerous GEK-bathythermograph sections at right angles to the mean direction of the surface current (040°). The plan called for *Crawford* to make an oceanographic section across each of the five locations at which *Atlantis* released floats. However, a series of untoward circumstances prevented *Crawford* from completing the first three sections. Of the two successful oceanographic sections, Section 1 was completed just prior to the passage of hurricane Ella near the work area. Both ships retired to Norfolk and Section 2 and the last group of floats were begun six days later. This delay should be kept in mind when considering the float tracks at the last two locations. The work was begun at the southwestern positions and proceeded to the northeast.

DIRECT CURRENT MEASUREMENTS

Two or more floats were released on each of five proposed sections; only those which were followed for a sufficient length of time to give a good estimate of velocity are shown in Fig. 1. Thirteen floats were used altogether; their statistics are summarized in Table 1. Navigational control by Loran *A* in this region is good to excellent for the purpose of tracking floats, but not sufficiently precise to provide an independent estimate of the depth of the floats, using the method described by SWALLOW and WORTHINGTON (1961). On only one occasion was it possible to determine the depth of a float above the bottom by observing both the direct and the bottom reflected signals. Previous experience suggests that the actual depth of the floats was within 10 per cent of the nominal depth. No attempt has been made to assess quantitatively the accuracy of the velocity measurements, but previous experience again suggests that the directions and speeds given are correct to $\pm$ 5 degrees and $\pm$ 1 cm/sec.

During the first phase of the cruise the floats showed a consistent pattern : the inshore ones moved southwestward and the offshore ones moved east or northeast. An unexpected result was the southward movement of float # 3 at a depth of only 800 m. Surface currents near this position averaged two to three knots toward the northeast according to the GEK. Obviously, at the time of observation the reference (zero axial velocity component) surface must have been shallower than 800 m. SWALLOW and WORTHINGTON (1961), slightly to the south of the present observation, had indicated that the zero velocity surface might rise as high as 250 m but their calculation was not supported by a direct measurement. At that time the shallow inshore countercurrent was thought to be detached from the deeper offshore current, but in the present study the countercurrent appeared to be continuous along the steep slope (approx. 1 in 10) of the continental escarpment from depths of a few hundred meters to at least 2500 m. However, there may have existed another band of southward moving water even further offshore where direct measurements were not made.

Both ships returned to the area after the passage of hurricane Ella, but time remained for only one more oceanographic section. Here the direct current measurements were extended in the offshore direction slightly beyond the range of the earlier observations. The eastern most float (#13) in this group moved southwest

Table 1. Summary of direct current observations using neutrally-buoyant floats

Float	Nominal depth (m)	Launch date	Launch Position Lat.	Launch Position Long.	Tracking time (hr)	Mean Velocity cm/sec.	Mean Velocity direction	Remarks
1	1200	4 X 62	33° 58′	75° 49′	4·0	19·0	230°	Lost after two fixes
2	1000	5 X 62	34° 04′	75° 46′	22·0	13·1	190°	Depth by bottom refl. = 1045 m.
3	800	6 X 62	34° 29′	75° 38′	38·0	21·1	205°–180°	Slight cyclonic curve
4	2000	6 X 62	34° 23′	75° 29′	49·0	5·5	065°	
5	750	8 X 62	34° 44′	75° 27′	—	—	—	Lost after one fix
6	2000	8 X 62	34° 39′	75° 18′	53·0	12·0	220°	
7	2000	9 X 62	34° 32′	75° 06′	32·0	5·5	090°	
8	2000	11 X 62	34° 56′	75° 02′	78·0	6·7	220°	Slowed from 8·6 to 3·8 cm/sec.
9	2300	12 X 62	34° 54′	74° 54′	48·0	10·1	055°	
10	1500	22 X 62	35° 14′	74° 50′	31·5	15·2	245°–205°	Cyclonic curvature
11	2000	22 X 62	35° 10′	74° 43′	27·0	17·7	220°–248°	Converging with #10
12	2500	23 X 62	35° 03′	74° 31′	3·2	20·0	185°	Lost after two fixes
13	2500	24 X 62	34° 58′	74° 34′	28·0	6·2	235°	

Table 2. Summary of photographic current meter measurements approximately 25 cm above bottom

Station	Depth (m)	Lat.	Long.	Date	Duration (min)	Velocity (cm/sec)	Velocity towards	Time for mud to clear
1	1300	34° 22′	75° 41′	7 X 62	7	*	200°–220°	50 sec
2	1100	34° 42′	75° 26′	9 X 62	47	4–8†	195°–210°	35 sec
3	3025	34° 58′	74° 29′	25 X 62	27	*	?	3 min

*Rotor of meter at rest during entire observation—estimated stalling speed = 4 cm/sec.
†Observed for about 4 minutes. Stalled for remainder of time.

at 6 cm/sec. Perhaps the deep current pattern which had prevailed during the earlier measurements was altered, but it is not possible to tell whether the anomalous behaviour of float #13 was an effect of the time lapse or of its more northeasterly location. GEK measurements indicated that the lateral position of the surface Gulf Stream was practically unchanged throughout the entire operation. Maximum surface velocities of two to three knots were consistently recorded about 10 km west of the line marked *A*.

Attempts to measure near bottom currents with the photographic current meter were only partially successful. Two lowerings showed evidence of a southgoing current with velocities too slow to turn the rotor of the meter. For one short interval when the rotor turned freely a velocity between four and eight cm/sec towards 195° was recorded. The depth of water was about 1100 m. None of the photographs showed ripple marks or other evidence of swift bottom currents. The

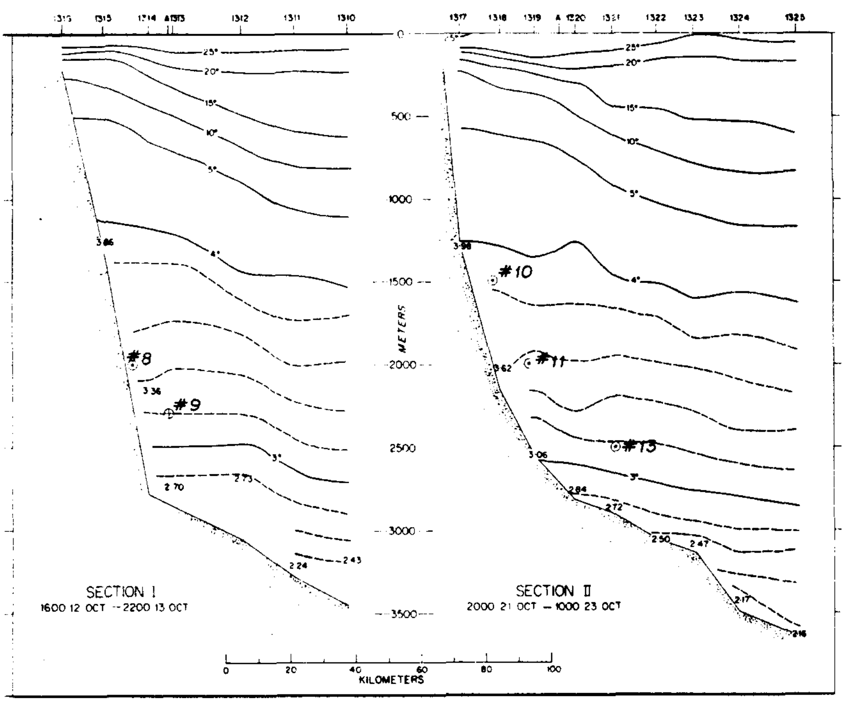

Fig. 2. Temperature profiles showing subsurface float positions as projected on the oceanographic sections. Vertical exaggeration = 50 ×.

current meter frame upon touching bottom generally created a large sediment cloud and the time required for its dispersal provided a rough measure of the current velocities. This cloud required from 35 sec to over 3 min for complete dispersal (Table 2) but this variation was certainly influenced by differences in the nature of the sediment involved.

WATER MASS ANALYSIS

The temperature sections (Fig. 2) show the location of the floats under the zone of maximum slope of the principal thermocline. The point marked *A* corresponds to line *A* on the horizontal chart as previously described.

The very precise relationship between potential temperature and salinity in most deep ocean waters which has emerged with the development and use of various salinometers in recent years has made increasingly useful the method of water-mass analysis based on calculation of salinity anomaly (HELLAND-HANSEN and NANSEN, 1926). The distribution of salinity anomalies in the range of potential temperature 3·51° to 4·00° (Fig. 3, from WORTHINGTON and METCALF, 1961) shows one feature of great interest to the present discussion : namely, the large body of fresh water with anomaly of − 5 parts per 100,000 or less centered about latitude 50°N in the Labrador Basin and extending southwest along the North American coast almost to latitude 30°N. This water mass, known as Arctic Intermediate Water (SVERDRUP, *et al.*

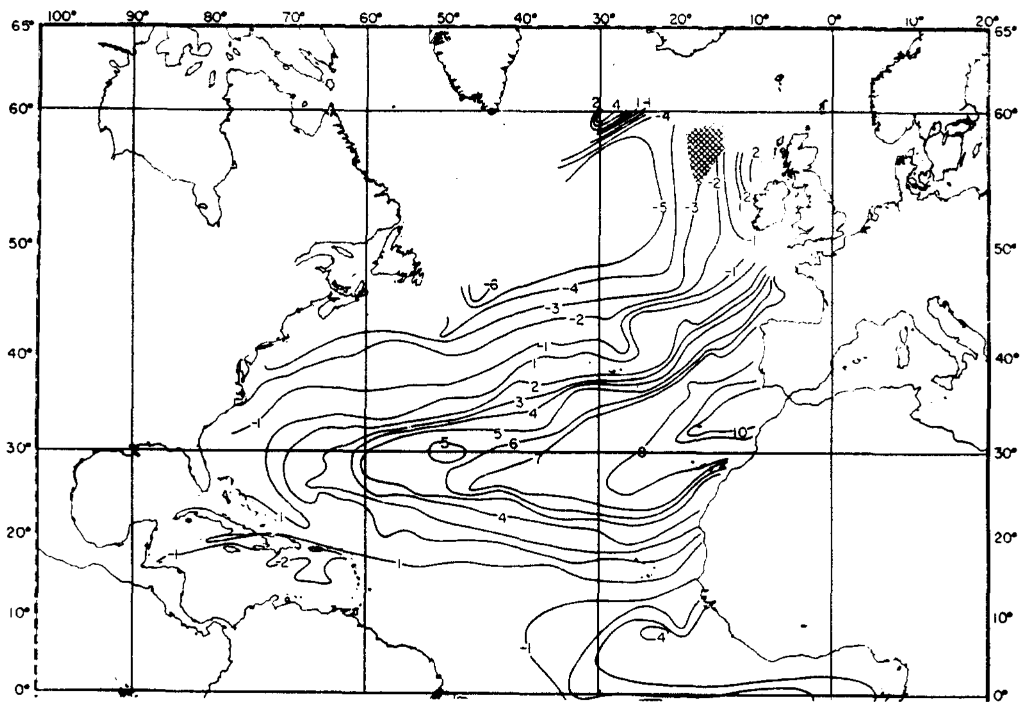

Fig. 3. Salinity anomaly in parts per 100,000 in the range of potential temperature 3·51° to 4·00°C. This is somewhat shallower than the depth of Wüst's upper intermediate oxygen maximum core. Reproduced from WORTHINGTON and METCALF (1961).

1941, p. 670), was previously thought to exert little influence outside its source region. But Fig. 3 plainly shows that the effect of this water type may be observed throughout the Slope Water Region (at depths of 1000 m to 1600 m), where salinity anomalies of − 1 to − 3 are commonly found. Negative salinity anomalies in this temperature range have also been observed consistently somewhat to the south of Cape Hatteras.

In the present operation a continuously recording *in-situ* salinometer was used in conjunction with the standard Nansen bottle technique. This is a great advantage where it is desirable to locate accurately salinity maxima or minima which might otherwise be observed only by fortuitous (or very close) spacing of sampling bottles. Salinity anomalies (from the mean θ–S curve of Worthington and Metcalf) near the five floats on Sections 1 and 2 were between -1 and -2 parts per 100,000 but on both sections minima less than -3 were observed at depths just below 1000 m (on stations 1315, 1319 and 1320). This corresponds to a potential temperature of about 4° and indicates a much greater southward penetration of Arctic Intermediate Water than even Fig. 3 suggests.

Salinity anomalies do not serve as an indicator of northern water in the range of potential temperature between 2° and 3°C. Within this range the Labrador Basin water closely adheres to the standard θ–S curve of the western Atlantic.

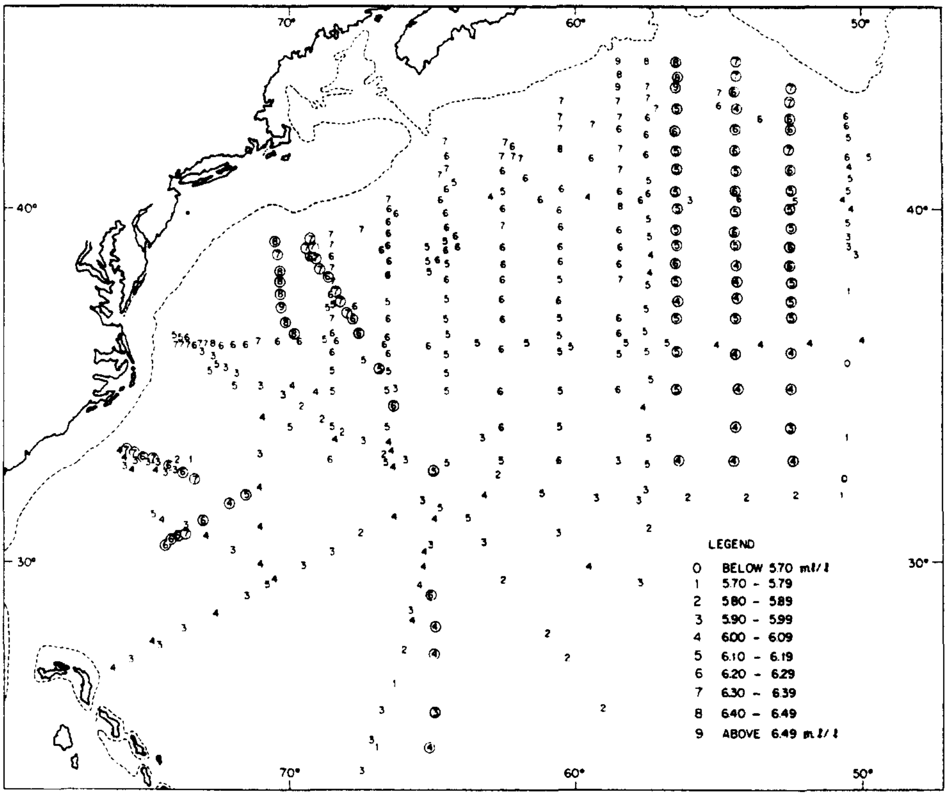

Fig. 4. Dissolved oxygen values at potential temperature 2·5°C. This corresponds roughly to a depth of 2700 m in the Slope Water Region and 3200 m in the Sargasso Sea, and is slightly shallower than the core of the lower oxygen maximum used by Wüst (1936) to trace the southward spread of the North Atlantic Deep Water. Circled numbers represent observations for which the chemical standardization was checked by an oxygen equilibrator.

However, in the Labrador Basin and deep slope water, values of dissolved oxygen are significantly higher than at corresponding levels in the Sargasso Sea. This is apparent not only at depths where the influence of the Arctic Intermediate Water is observed but also at 2·5° potential temperature as shown in Fig. 4. High oxygen values south of Cape Hatteras serve as indicators of the spreading of this slope water to the south, just as the salinity anomaly distribution indicated a southward extension of the adjacent intermediate water.

A majority of the observations shown in Fig. 4 are from the IGY and from "Gulf Stream '60" (Fuglister, 1963). Because of changes in the method of oxygen determination at W.H.O.I. and refinements in technique (CARRITT, 1959), there are undoubtedly included here many observations of greater precision than accuracy. That is to say, it is quite likely that the values of oxygen reported in the past may have been consistently low or (less often) high throughout an entire cruise. For this reason no attempt has been made to contour oxygen values in Fig. 4. But, in spite of this difficulty, there emerges a pattern which is indicative of the spreading of high oxygen water southward past Cape Hatteras. This is especially true when only those values based on the dependable equilibrator technique (CARRITT, 1963) are considered.

It is apparent that this southward flow of relatively oxygen-rich water into the Sargasso Sea from the Slope Water Region provides a means by which the oxygen level of the deeper portions of the Sargasso Sea may be maintained. How this effect may compare in importance with that of the large cyclonic meanders which break off to the south of the Gulf Stream further east, as a factor in the cross-stream transfer of properties, is not known.

Figs. 3 and 4 reveal another clue concerning the nature of the deep ocean circulation. WÜST (1936) and STOMMEL (1957) both envisaged the undercurrent as a continuous feature along the entire western side of the Atlantic, so it is natural to attempt to trace its further progress southward using the properties already discussed. In neither figure is the southward spread of northern water as pronounced as in WÜST's (1936; 1963) analyses using the *kernschicht* method. In the salinity anomaly distribution (Fig. 3) there is an indication of the spread of relatively fresh water along the north side of the Antillean arc as far eastward as Puerto Rico, but this effect is obscured by the very strong salt core extending outwards from the Mediterranean Sea. At the 2·5° potential temperature level, the oxygen distribution (Fig. 4) is even less indicative of southward flow, although recent observations (by the author) north of Puerto Rico averaged 6·10 to 6·19 ml/l in this temperature range. The fact that Fig. 4 fails to indicate the progress of the undercurrent south of 30°N may be due entirely to the scarcity of post-IGY data between 24°N and 30°N, but it is suggested that the salient submarine ridge extending southeast from Cape Fear tends to divert the deeper portion of the undercurrent away from the continental margin, and that if a deep southward flow exists to the east of the lesser Antilles, it is not likely to be a direct continuation of the flow observed at Cape Hatteras. This question, and also that of whether a continuous current can be identified in the range of potential temperature 3·5° to 4·0°C, as the distributions of salinity and oxygen suggest, must be determined by further direct current measurements.

DYNAMIC CALCULATIONS

Geostrophic calculations performed in the standard way (e.g. LA FOND, 1951) are supposed to describe fairly closely conditions actually encountered in swift narrow currents. WEBSTER (1962) has demonstrated that the calculated geostrophic velocity may be expected to depart by about 10 per cent of the value of the observed velocity. Other estimates of non-geostrophic departures likely to be encountered range generally between 5 per cent and 25 per cent of the computed velocity, (VON ARX, 1962, p. 254), but WORTHINGTON (1954) has suggested that calculations of this sort cannot be relied on in mid-ocean away from the swift boundary currents.

The question of the reliability of dynamic calculations in general is inseparable from that of the determination of a suitable reference surface by which to transform relative into absolute velocity. With the recent improvements in direct current measuring devices (particularly the neutrally-buoyant float) it might seem that the reference level problem no longer exists; in point of fact, the situation is improved but far from settled. Thus, in addition to the usual assumptions concerning friction, accelerations and simultaneous observation inherent in the geostrophic method, any attempt to transform relative velocity into actual velocity by means of direct current measurements involves an additional critical assumption : namely, that there is an equivalence between the velocity calculated from a trajectory (or observed

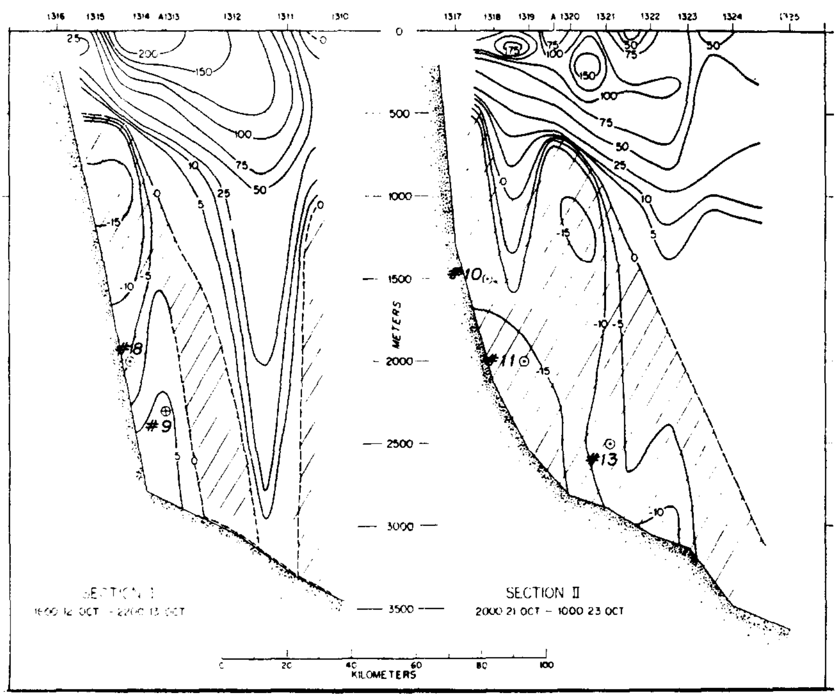

Fig. 5. Velocity profiles (cm/sec) computed geostrophically and conforming to direct current measurements. The dotted portion of the zero velocity line represents extrapolation (see text). Light shading indicates southward flow. Isotachs were drawn by considering the computed velocities between stations at each standard depth to be point values, but as the average station spacing on both sections is less than 15 km this method is quite adequate when compared to the more rigorous method of WÜST (1924).

at a fixed point, in the case of an anchored current meter) and the actual mean velocity at the same depth between a pair of oceanographic stations. Since this assumption concerns horizontal variations in the pressure field and the corresponding velocity variations between observations, best results will be secured from close spacing of oceanographic stations and high density of direct observations, coupled with some prior knowledge of the scales of variability likely to be encountered in a given area. Direct observations are difficult and costly so they are still not available in great numbers.

In the present instance the reference level was established in the following manner. Where a direct current measurement was available on or near the oceanographic section the computed relative velocity was adjusted to agree with the observed velocity component normal to the section. In Section 2 an interpolated velocity between floats #11 and #13 was supplied at the appropriate depth for station pairs 1319–1320 and 1320–1321. At the ends of the sections where direct current measurements were lacking, the deduced zero velocity level was extrapolated in a manner consistent with the salinity observations and also taking into account the pattern of direct current observations at all locations. Thus, it seemed reasonable to continue the zero level at 500 m on the inshore end of Section 1. For the offshore portion of Section 1, velocity at the bottom was assumed to be zero* and on Section 2 the zero level was arbitrarily extrapolated as indicated by the dotted line in Fig. 5. It would have been equally possible to have assumed zero velocity at the bottom here also, but since the calculated vertical shear of velocity in this portion of the profile is small no large change in the computed volume transport would result (Table 3).

Table 3. *Computed volume transports* (Million m^3/sec)

Section	*Northward above Reference level*	*Southward below Reference level*	*Net Northward*
1	64·3	3·9	60·4
2	53·6	11·7	41·9
2 assume zero velocity at the bottom from Stn. 1322 eastward			
	59·7	10·5	49·2

This method of extending the reference level departs considerably from that employed by SWALLOW and WORTHINGTON. In 1958 the potential temperature of the computed zero velocity level was observed to be fairly constant at 3·5° $\pm$ 0·1°C. Consequently, the zero level was extrapolated along surfaces of constant potential temperature. In the present instance the direct current measurements are somewhat spread out across the undercurrent and possibly for this reason the recurrence of the computed zero velocity level at a particular temperature level was not observed.

The estimated values of the volume transport of the undercurrent are not greatly different from the previous estimate (SWALLOW and WORTHINGTON, *op. cit.*) of 6·7 million m^3/sec. But the possibility of relatively large changes in volume transport taking place in a short time interval is inherent in the present computations. Southgoing velocities observed on the present cruise were in the range of 7–21 cm/sec; those reported in 1961 were between 9 and 18 cm/sec. It is clear that large changes in velocity are not required to explain the variations in calculated volume transport.

The fact that the deeper of the two floats on Section 1 was moving northeastward, although located in water with a slight negative salinity anomaly, deserves mention. Obviously it is not possible meticulously to relate velocity (direction) and salinity anomaly, but in general there does seem to be a reasonable correlation. This particular float (# 9) was located in a small horizontal temperature (and density)

*The mid-depth zero velocity surfaces indicated in the eastern portion of Section 1 result from the nature of the individual calculated velocity–depth traces and the assumption of zero velocity at the bottom. They are probably not meaningful; except between stations 1311 and 1312 there was little change of computed velocity with depth in the deeper portion of Section 1. The assumption of a mid-depth zero level between stations 1311 and 1312 would, of course, have resulted in a southward flow below the zero level.

gradient. Possibly the northeasterly direction of this float indicates an entrainment of a portion of the undercurrent into the main (northeastward) flow. Unfortunately, no direct observations were made in the eastern portion of Section 1 where the deep isotherms again slope downward in the offshore direction.

A noteworthy feature of the velocity profile computed for Section 2 is the sinusoidal shape of the zero velocity surface between stations 1318 and 1321. On the basis of the salinity observations it seemed that the dip in the zero-level between stations 1318 and 1319 might be unreal, but an attempt to raise the zero-level in that interval resulted in a velocity near the bottom much higher than any observed. The small area of high velocity just below 1000 m centered near station 1320 corresponds very closely to a core of maximum negative salinity anomaly, (— 0·035‰).

The individual velocity-depth traces for those station pairs whose reference level had been established by a direct current measurement (or by linear interpolation) were examined to see whether or not the zero-level fell in the layer of minimum change of relative velocity with depth as suggested by DEFANT, (1961, p. 494). In three cases out of seven the zero-level fell within the region of minimum shear; in the other four it was located very slightly above that region.

SUMMARY

A southward flowing western boundary undercurrent was located again. It seems likely that this current is a persistent feature of the ocean circulation, although it has been observed directly on only three occasions. A close scrutiny of available oceanographic data shows that deep water slightly fresher and richer in dissolved oxygen than similar water in the Sargasso Sea has frequently been observed south of Cape Hatteras. The flow, where it has been observed east and south of Cape Hatteras, occupies an area close to the steep face of the continental escarpment and the upper portion of the continental rise. The width of the deep flow is variable; probably between 20 km and 80 km; it may occasionally be divided into two or more streaks (*cf.* SWALLOW and WORTHINGTON, 1961). Velocities greater than 25 cm/sec have not been observed in it. The depth of the zone of no motion may be as shallow as 500 m near the top of the continental escarpment, increases rapidly in the offshore direction, and must extend to the bottom, at least in those cases where the offshore float was moving to the northeast. The volume transport of the undercurrent is estimated to be between 4 and 12 million m^3/sec which is considerably less than that suggested by STOMMEL (1957) or observed by VOLKMANN (1963) in the westward flow in the Slope Water Region (30 and 40 million m^3/sec, respectively).

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