Description of the Circulation on the Continental Shelf

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## CHAPTER 4

## A DESCRIPTION OF THE CIRCULATION ON THE CONTINENTAL SHELF OF THE EAST COAST OF THE UNITED STATES\*

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Abstract. The circulation on the continental shelf of the east coast of the United States is discussed, including the historical development of the concepts, and the surface and bottom circulation based on drift-bottle data and sea-bed drifter data. Suggestions for future research to further our understanding of the circulation problem are offered.

## INTRODUCTION

The intent of this report is to provide an interpretation of the large amount of drift-bottle and sea-bed drifter data acquired during the 1960s over the continental shelf of the east coast of the United States and to suggest some approaches to elicit a better understanding of the circulation problem in the future.

Inasmuch as the literature about the continental shelf is copious and scattered, it appears warranted to this author to review the information we now have to provide a firm basis for designing future research programs.

## THE CONTINENTAL SHELF

The bathymetry of the area discussed here is handsomely illustrated in UCHUPI's (1965) chart. It comprises (Fig. 1):

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The outer edge of the shelf north of Hatteras is cut by numerous canyons, one of which, the Hudson Canyon, can be traced across the shelf to the approaches to New York harbor.

North Atlantic slope water impinges on the outer edge of the shelf north of Cape Hatteras whereas the Florida Current sweeps northerly along the outer edge of the South Atlantic



FIG. 2. Typical temperature (T), salinity (S), density ( $\sigma_t$ ) sections across the continental shelf.

A. After COLTON et al., (1968), section E across Georges Bank, March (left) and September (right) 1966.

- B. After KETCHUM and CORWIN (1964), section south of Montauk Point, February (left) and July (right) 1957.
- C. After STEFANSSON, ATKINSON and BUMPUS (1971), section 3 across Raleigh Bay, February (left) and June (right) 1966.

Bight. Several typical temperature, salinity, density sections across these areas are shown in Fig. 2. Rotary tidal currents flow over the shallow parts of Georges Bank at approximately 50–100 cm/sec, causing vigorous mixing of the water column and permitting the development of a weak seasonal thermocline for a period of only 2 months during the mid-summer. Tidal currents elsewhere over the continental shelf are much slower (ANONYMOUS, 1971; table 5) and have little effect on the vertical structure of the water column.

Fresh water is contributed via sounds, bays and river mouths, from the intricate river systems draining the eastern hinterland of the United States, to the inshore edge of the shelf. According to BUE (1970) the average stream flow into the Atlantic coastal waters amounts to  $355,000 \text{ ft}^3/\text{sec}$  (9944 m<sup>3</sup>/sec), roughly 20% of the average flow out of the coterminous United States, but twice the national average on the basis of cubic feet per second per square mile. The annual flows emptying into the several segments of the continental shelf and comparative information appear in Table 1.

## HISTORICAL DEVELOPMENT OF CIRCULATION CONCEPTS

Investigations into the circulation of the waters of the continental shelf of the east coast of the United States have progressed more or less from north to south, conducted for the most part by biologists. At one time it was commonly thought that the Labrador Current flowed from the Grand Banks westward past Nova Scotia and so southward as far as Florida (SUMNER, OSBURN and COLE, 1913). This assumption was developed on the basis of temperature, inasmuch as the coastal-water temperatures are colder than those offshore (see SCHROEDER, 1966; WALFORD and WICKLUND, 1968; COLTON, MARCK, NICKERSON and STODDARD, 1968; COLTON and STODDARD, 1972). Charts of salinity at various levels in the Atlantic Ocean (presently being prepared by L. V. Worthington) do suggest the presence of Labrador water in the slope water and over the outer edge of the continental shelf as far as Cape Hatteras, and penetration into the Gulf of Maine through Northeast Channel. The composition of this slope water may be modified from time to time, with additional increments of Labrador coastal water, resulting from movements of the Icelandic low-pressure area in winter (WORTHINGTON, 1964).

## Gulf of Maine

Observations and conjecture about the circulation of the Gulf of Maine were made prior to Bigelow's classic studies (BIGELOW, 1927) which commenced in 1912, but no serious effort to study the Gulf of Maine as a whole occurred prior to this period.

SCHOPF (1968) reported

Surprisingly little was known of surface or bottom circulation in the Gulf of Maine and connected bodies of water in Verrill's time. As late as 1881, Mitchell was unraveling problems of tidal drift, and there was almost no information on non-tidal drift. Verrill (1873) mentioned that there was a southward flowing cold current on the shelf south of New England and landward of the Gulf Stream, but he erroneously thought this derived directly from a cold current flowing southwestward along the seaward side of Nova Scotia. Drift bottles were first systematically used in circulation studies in 1919.

BIGELOW (1927) recognized that "water from the northeast (low in temperature) does flow past Cape Sable into the Gulf of Maine for a time in spring, sometimes in summer". This

TREAM FLOW INTO THE THREE MAJOR SEGMENTS OF THE ATLANTIC COAST BETWEEN THE ST. CROIX RIVER AND	CAPE SABLE, FLORIDA, 1931–1960
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TABLE	

	Stream flow (m <sup>3</sup> /sec)	Length of coastline (km)	Stream flow per length of coastline (m <sup>3</sup> sec <sup>-1</sup> km <sup>-1</sup> )	Stream flow (10 <sup>9</sup> m <sup>3</sup> /yr)	Area of shelf to (100 m isobath) (10° m <sup>2</sup> )	Stream flow per unit area of shelf (m/yr)
Gulf of Maine	1676.6	735	2.28	52.8	152.6	*0.35
Middle Atlantic Bight	4977.8	1060	4.70	156.8	97.5	1.61
South Atlantic Bight	3290.0	1510	2.18	103.6	100.4	1.03

\*BIGELOW (1927) estimated 0.79 m/yr for the whole Gulf of Maine.

Stream flow data from Bug (1970). Note that the stream flow/length of coastline for the Middle Atlantic Bight is more than two times that for the Gulf of Maine and the South Atlantic Bight; the stream flow/area of shelf for the Gulf of Maine is one-third that for the South Atlantic Bight and  $\simeq$  one-fifth that for the Middle Atlantic Bight.

water originates in part from the Gulf of St. Lawrence and partly from Banquereau and the Sable Island Banks. It is not Labrador water. BIGELOW (1927) developed his concepts of the circulation of the Gulf on the basis of the distribution of temperature and salinity, dynamic computation, analysis of current-meter data, the return of drift bottles released along strategic sections, and the movement of fish eggs and larvae. He described the seasonal development of a great Gulf of Maine eddy moving counterclockwise around the basin of the Gulf. It varied in velocity and detail from season to season and was complicated by subsidiary eddies. The southern limb of the Maine eddy coincided with the north side of Georges Bank over which a clockwise eddy developed. There was some spilling of water out of the Maine gyre south past Cape Cod. The velocity of the net drift about the margin of the Gulf was on the average 7 nautical miles per day. He also recognized an indraft of slope water through the Northeast Channel between Browns and Georges Bank, concluding that the variations in the temperature and salinity of the deep water of the Gulf from time to time are due to fluctuations in the volume of the inflowing slope water. COLTON (1968, 1969), however, has demonstrated that the temperature-salinity characteristics of the subsurface waters adjacent to the entrance to Northeast Channel change slightly from time to time. Variations in the composition as well as the volume of the slope water indraft through Northeast Channel are reflected in the warming or cooling trends of subsurface water within the Gulf of Maine.

The large amplitude of the tidal rise and fall and the strong tidal currents were considered to be the major source of energy for the residual drift, only slightly modified by the wind. HUNTSMAN (1924) reported

It has recently been found that the coastal waters inside the continental shelf are in slow circulation at rates ranging for the most part from 2 to 6 miles per day. The course of the horizontal circulation is determined largely by the configuration of the bottom, the water rotating in a contra-clockwise direction around basins, and in a clockwise direction around islands and shoals.

He reasoned that the deflective effect of the earth's rotation on the pumping action of the tides acted as an "imperfect valve", diverting the flood and ebb currents to the right. Thus, while the tide flows into a confined basin, it is prevented by the right-hand shore from turning to the right; hence it must rise higher against the shore or flow farther parallel to the coast than it would were it not confined. The ebb current is similarly pressed against the opposite shore.

On the basis of drift-bottle data and the drift of haddock larvae, WALFORD (1938) concluded that Georges Bank supplies its own stock of haddock by retaining a sufficient portion of the eggs and larvae in an eddy on the Bank. He noted that at the end of April 1931 a current was carrying the eggs produced on the Northeast Peak toward Nantucket Shoals. Between late April and May there was a change; part of the current continued west toward Nantucket Shoals, and an "inner current turned northerly and easterly around Georges Shoals, thus conserving the supply of larvae". On the other hand, he noted losses of haddock eggs in April 1932 from off the northern edge and from the southwest portion of the Bank. Presumably a large amount of Bank water was carried off into the slope water zone. CHASE (1955) developed an empirical formula for predicting the brood strength of Georges Bank haddock on the basis of the lengths and strengths of the offshore winds (northwest) after the time of minimum temperature.

The observed drift of a population of the pteropod *Limacina retroversa* drawn into the Gulf of Maine from the east and the position of water of salinity greater than 33‰ at 50 m led REDFIELD (1939) to conclude there was a drift of subsurface waters around the Gulf

during the winter-spring through a distance of 150 miles in 4 months, at a rate of about 1.25 miles per day. After examining the distribution of several populations of chaetognaths, REDFIELD and BEALE (1940) concluded that the permanent occurrence of the only truly endemic chaetognath, Sagitta elegans, depended on a relatively stable eddy on Georges Bank. The presence of the other species (Eukrohnia hamata, Sagitta maxima and S. lyra) was dependent on their being carried into the Gulf via the Northeast Channel by deep currents and S. serratodentata by surface currents from the east. REDFIELD (1941) reported the Maine eddy as being open to the southeast over eastern Georges Bank and Northeast Channel in autumn and winter, but becoming a closed recurrent eddy in late spring, i.e. a movement from the southern portion of the Gulf toward the northeast quarter.

In a further study of the distribution and reproduction of Sagitta elegans on Georges Bank, CLARKE, PIERCE and BUMPUS (1943) noted

It is of special interest that a nucleus of these species apparently retained its position on the Bank throughout the winter, during the period when the breakdown of stratification in the surrounding areas might be expected to make possible a flow of Gulf water directly across the Bank. The bubble of mixed water on the Bank therefore either fails to be dislodged by hydrographic forces, or is renewed so slowly that the population of Sagitta is able to maintain itself within the mixed area despite the water movement.

The completion of an analysis by HAIGHT (1942) of current measurements with tide poles at lightships and a few additional locations gathered over a 30-year period along the Atlantic coast gave evidence of the extreme variability of the non-tidal current. Haight reduced the data to tables and charts of tidal currents relative to the Greenwich transit of the moon, the non-tidal currents by month, and wind currents relative to wind direction. The non-tidal data for Portland and Boston Lightship are of relatively short duration, 8 months during the latter part of 2 years at Portland and 16 months at Boston. The residual drifts are widely disparate in direction from month to month, ranging from 047°T to 254°T (clockwise) for Portland and 012°T through 357°T for Boston, with speeds ranging from 0.02 to 0.28 knot and from nil to 0.10 knot respectively. Where there are data for a given month in more than 1 year at these stations, the direction and speed are not consistent. No useful trends appear available from these data for the Gulf of Maine.

Drift-bottle data obtained in 1931, 1932, 1933, 1934, 1953, 1955 and 1956 were carefully analyzed by DAY (1958). He attempted to determine the influence of the wind on the surface circulation. He concluded:

Bigelow's July-August pattern, a relatively closed circulation and hence highly conservative, seems to evolve with the seasons, possibly as a result of vernal warming and the consequent formation of density structure. In February and March the Gulf eddy is ill-defined on the south as water from Area I (Massachusetts Bay) moves southward out South Channel. The Georges eddy is not apparent, the net movement of surface water being southwest. By late April, the two counter rotating eddies are taking form with the appearance of a northeasterly movement along the northern edge of Georges Bank where the two eddies are confluent.

At any time during the seasons considered (late February through June), the prevailing circulation may be strongly modified by winds. Offshore winds from the northwest quadrant have a pronounced effect on the Georges Bank region, tending to destroy the eddy. Northeasterly winds, even of short duration, speed up the northern segment of the Gulf eddy and increase the flow out South Channel. Winds from the west and southwest seem to give a strong onshore component to the waters south of Cape Cod and to accentuate the northward flow from Georges Bank into the Bay of Fundy.

In the region of Georges Bank, a marked difference between conditions in the 1930s and the 1950s is apparent. With the exception of the April cruise of 1932, when the winds were uniformly offshore, the Georges eddy during the 1930s was far more pronounced during the last few years. Though there are indications of an eddy in 1955, it is markedly absent in 1953 and 1956.

#### D. F. BUMPUS

Some of the previous interest in the circulation of the Gulf of Maine has been concerned with the advection of water off Georges Bank to the south and west into the slope water zone. In a study of the source of water contributed to the Bay of Fundy by surface circulation, BUMPUS (1960) reported

The area from which drift bottles at the surface enter the Bay of Fundy is restricted during January to the immediate approaches to the Bay. It gradually expands during the spring to a maximum in early summer, encompassing most of the Gulf of Maine and part of Georges Bank as the result of the confluence of the Georges and Maine eddies. The source area gradually retracts during the autumn to the eastern side of the Gulf of Maine and northward away from Georges Bank.

BUMPUS (1960) also noted that after a period of drought the effect of high runoff on the circulation around the north and west sides of the Gulf produced more consistent and faster drift.

John Colton spent many years in the Gulf of Maine and on Georges Bank studying the icthyoplankton and the hydrography. He was particularly conscious of the band of relatively warm slope water along the southern edge of the Bank. He noted the contrasting tongues of Bank and slope water where tropical and boreal plankton mix and drift slowly out of the area. COLTON (1959) observed in May 1956 that the slope water extended further than usual over the southeast corner of the Bank. Juvenile stages of a number of tropical and oceanic species were captured in the plankton tows, whereas the endemic species were found to be in various stages of decomposition.

The presence of eyed flounder (*Bothus ocellatus*) postlarvae and myctophid postlarvae well into South Channel in late September and early October 1956 (COLTON, 1961) and many species of oceanic copepods in the southern Gulf of Maine–Georges Bank area during the autumn of 1956 and winter of 1957 (COLTON, TEMPLE and HONEY, 1962) suggest an overflow of oceanic water which they believed occurred in July or August. Confirmation of this sort of event was observed 10 years later when the ESSA ship *Explorer* followed the 15° isotherm at 200 m (center of the Gulf Stream) from Cape Hatteras eastward toward the Grand Banks on a monthly schedule between September 1965 and October 1966. HANSEN (1970) noted: "Between the cruises of May and June, 1966, an abrupt change in the nature of the meandering of the stream occurred. A large loop in the thermal front pushed up into very shallow water on Georges Bank, the 15° isothermal surface, in fact, intersected the continental shelf."

Through the use of drift bottles and transponding buoys (WALDEN, KETCHUM and FRANTZ, 1957) COLTON and TEMPLE (1961) observed that with the exception of mid-summer when the Georges eddy is most pronounced and southerly winds predominate, the surface drift is offshore in the direction of the slope water band. The currents below the surface move at a slower rate than at the surface but in a similar direction.

BRYANTSEV (1965) noted that the wind shifts the borders of the Georges Bank and Labrador water masses along the 50- to 100-m isobaths in summer, substantially influencing the efficiency of the herring fishing effort. Catches were increased at the time of southwest winds and decreased during northeast winds.

Other experiments with transponding drift buoys (BUMPUS, 1958; Figs. 21, 22 and 23) conducted in May–June, October, December–January over various parts of Georges Bank and the Gulf of Maine confirm the presence of the eddies and the drifts off the bank into deep water.

By 1962 sufficient drift-bottle data had been accumulated to produce an atlas of monthly charts showing the inferred speed and direction of surface drift (BUMPUS and LAUZIER, 1965). They noted:

The indraft off Cape Sable, from across Browns Bank and the eastern Gulf of Maine into the Bay of Fundy, is the chief characteristic during the winter season. A southerly flow develops along the western side of the Gulf of Maine and continues past Cape Cod through Great South Channel. Between the indraft into the Bay and the southerly flow along the western side of the shelf several irregular eddies develop by February. An area of divergence north of Georges is well developed by February.

The Gulf of Maine eddy develops rapidly during the spring months so that one large cyclonic gyre encompasses the whole of the Gulf of Maine by the end of May. There is an indraft on the eastern side of the gyre from the Scotian shelf and Browns Bank. Abreast of Lurcher Shoal the drift may continue on northward into the Bay of Fundy, or it may turn westward toward the coast of southern Maine, continue southward across Massachusetts Bay, where it may divert into Cape Cod Bay or turn east of Georges Bank.

The Maine eddy, which reached its climax in May, begins to slow down in June. By autumn and winter the southern side breaks down into a drift across Georges Bank.

The few returned drift bottles from winter releases on Georges Bank suggest a southerly flow across the area during the winter months, with a westerly component across Great South Channel. During the spring months an anticyclonic eddy develops on Georges Bank. The northerly side of this Georges eddy is common with the southern side of the Maine eddy; the area of divergence continues along the northern edge of Georges Bank. A persistent westerly drift along the southern side of the Bank continues across Great South Channel.

During the summer the eastern side of the Georges eddy veers southerly and offshore. With the onset of autumn the west side of the Georges eddy breaks down into a westerly and southerly drift.

Employing sea-bed drifters (LEE, BUMPUS and LAUZIER, 1965) to examine the residual drift near the bottom, LAUZIER (1967) reported

The shelf area of the eastern Gulf of Maine, from Browns Bank to the entrance of the Bay of Fundy is considered as an area of large-scale convergence where a shoreward bottom drift is associated with an offshore surface drift, indicating upwelling along the coast as far north as St. Mary's Bay.

A zone of divergence is located in the deep waters between Georges Bank and Browns Bank.

A two-way bottom drift, similar to the surface drift, at the entrance to the Bay of Fundy is indicated but the demarkation is more definite on the bottom than at the surface.

He estimated the rates of drift to be between 0.2 and 0.7 nautical miles per day.

In examining the surface and bottom drift along the coast of Maine from Machias Bay to Cape Ann to a distance of 29 km offshore, GRAHAM (1970) found the most important feature of the circulation along the coast was upwelling. Surface water moved parallel to or offshore from the coast with a compensatory movement inshore along the bottom. The studies of SCHLEE *et al.* (1971) appear to confirm this.

#### Middle Atlantic Bight

The early investigators in the Middle Atlantic Bight were chiefly concerned with the distribution of various fauna and of temperature. They referred to the warmer waters of the slope water as the "Gulf Stream" lying about 85 nautical miles south of Martha's Vineyard and Nantucket. SUMNER, OSBURN and COLE (1913) noted: "But the influence of the Gulf Stream extends much nearer to the coast than the edge of the continental shelf.... The presence nearly every year in Vineyard Sound of considerable masses of Sargassum bacciferum, with its attendant fauna shows that the strong southerly winds may drive the surface water of the Gulf Stream as far as the mainland of Massachusetts".\* The catastrophic

<sup>\*</sup>It is not uncommon in August to find the Portuguese man-of-war, *Physalia physalis*, a species prevalent in the Sargasso Sea, off Nobska Beach, Woods Hole, yet we do not find the temperature and salinity of the water there consistent with that of the Gulf Stream or the Sargasso Sea. The *Physalia* drift there under the influence of the prevailing summer southwest winds.

mortality of tilefish on the outer edge of the shelf in March and April, 1882, was attributed to a direct flow of the cold inshore waters out over the area usually occupied by warmer slope water (BUMPUS, 1899).

Bigelow intended to describe the circulation in the Middle Atlantic Bight following his description of the Gulf of Maine, but succeeded only in describing the temperature and salinity regimes (BIGELOW, 1933; BIGELOW and SEARS, 1935) before he became engrossed with the description and distributions of fishes. The record of surface temperatures from lighthouses and lightships was reported for the periods 1881–1885 (RATHBUN, 1887) and 1928–1931 (PARR, 1933) in an examination of the seasonal changes in the temperature of the inshore waters from a geographic-ecological point of view. BUMPUS (1957a) reported all of the data for the period of record, 1873–1956, and STEARNS (1964, 1965) examined the data (1873–1961) for secular changes in the climatic warming and cooling of this region.

The current pole data of HAIGHT (1942) in the Middle Atlantic Bight area are most instructive. The non-tidal drifts at *Nantucket* Lightship ( $40^{\circ}37'N$ ,  $69^{\circ}37'W$ ) were toward the southeast quadrant from December through February, at less than 0.1 knot, toward the southwest in March at 0.3 knot, and toward the northwest quadrant at 0.1 to 0.18 knot during the remainder of the year. The Ambrose Channel ( $40^{\circ}27'N$ ,  $73^{\circ}49'W$ ) drifts were chiefly toward the southeast ranging from 0.04 to 0.22 knot, whereas those for *Scotland* Lightship ( $40^{\circ}26.6'N$ ,  $73^{\circ}55.2'W$ ) between Ambrose and Sandy Hook were southeasterly but slightly faster, up to 0.32 knot. The *Barnegat* Lightship ( $39^{\circ}45.8'N$ ,  $73^{\circ}65'W$ ) drifts were weak (0.01 to 0.10 knot) and toward the southeast to south except for a tendency during the late spring throughout summer toward the northeast. Similar drifts were exhibited for *Northeast End* ( $38^{\circ}57'N$ ,  $74^{\circ}29.6'W$ ), *Winter Quarter* ( $37^{\circ}55.4'N$   $74^{\circ}56.4'W$ ), and *Chesapeake* ( $36^{\circ}58.7'N$ ,  $75^{\circ}42.2'W$ ) Lightships where the southerly to southeasterly flow was interrupted by easterly–northeasterly drifts during the summer. The net drifts for Diamond Shoals were all northeasterly ranging between 0.24 and 0.83 knot.

We have the first real understanding of the circulation problem on the shelf when ISELIN (1939, 1940) pointed out the fundamental difference in the distribution of salinity between oceanic and coastal waters. In oceanic waters the salinity decreases with depth partly counteracting the stability resulting from the vertical temperature gradient, whereas over the continental shelf the salinity usually increases with depth. This led to his conclusion that "The coastal circulation has an offshore component at the surface and an inshore component beneath". He further noted: "The coastal waters, because of their relative freshness, are at most times of the year less dense than the corresponding layer offshore and consequently a current is maintained which, for some reason not clearly understood, tends to have its greatest strength just outside the 100-fathom curve." In most areas where there is sufficient runoff, the density distribution is the cause and not the result of currents. He also postulated considerable mixing and exchange with slope water across the 100-fathom line, continuous small-scale mixing processes, and sudden large offshore movements.

MILLER (1950) examined the mixing processes along the 100-fathom line south of Rhode Island in November 1948. He found major transfers taking place between surface slope water and the shelf water. He found non-isentropic mixing, requiring an internal energy source which he attributed to internal surf in the many-layered medium converging on the edge of the shelf.

In the course of determining the flushing of New York Bight with reference to the newly commenced discharge of acid-iron waste, KETCHUM, REDFIELD and AYERS (1951) found that the water circulation in this area is rapid, the flushing time for the area being 6 to 10 days in

spite of a nine-fold variation in the river flow. The sea water flux ranged from 47 to 224 times the volume of the river flow. The flushing rate is independent of the river flow, yet changing patterns were expected when the river effluent was small, and well-defined patterns developed when the effluent increased.

In a further report on the barged iron-acid discharge in the New York Bight, REDFIELD and WALFORD (1951) reported:

The circulation patterns are quite variable, ranging from the most common type, which indicates that the river effluent escapes in a narrow band along the New Jersey coast, to one in which the river effluent is distributed widely over the surface area. Although five of the six distributions observed can be accounted for by the associated oceanographic forces and the winds, so many variables contributed to the circulation that predictions are not feasible.

The fresh water from the river moves seaward across the area immediately off New York at an average rate of about 3 miles per day.

Following drift-bottle studies in the Rhode Island Sound and the New York Bight during the summer of 1951, POWERS and AYERS (1951) concluded

The evidence gained from bottles recovered locally and from those recovered in the Cape Hatteras area indicates the existence of two southerly coastal drifts, an inshore drift consisting mainly of a series of eddies and serving to transport the main portion of undetermined pattern and speed whose characteristics probably more nearly approach those of true shelf water than those of inshore waters.

SHONTING'S (1969) array of self-recording current meters covering a square kilometer in Rhode Island Sound during the mid-summer of 1967 demonstrated that the surface flow above the pycnocline was strongly isolated from the layer below. The semidiurnal tide was observed in both layers. A westsouthwest mean drift prevailed in the upper layer in contrast to a more random and rotary flow in the lower layer. He found that the mean motion kinetic energy and the turbulent kinetic energy of the upper layer were four to five times that of the bottom layer.

Dye experiments in the vicinity of *Ambrose* Lightship, the head of Hudson Canyon, and the adjacent New Jersey shore revealed that the circulation in this area during the summer and autumn is dominated by weak counterclockwise eddies close to the New Jersey shore which carry the low salinity Hudson-Raritan effluent southward. A northward-moving drift is encountered about 10 miles offshore in the more saline water (COSTIN, DAVIS, GERARD and KATZ, 1963).

GRISCOM and SOMMERS (1969) reported the results of their current-meter measurements at Texas Tower 4, 70 miles southeast of New York. The meter at a depth of 100 ft, where the water was 185 ft deep, made hourly measurements for 10 days in April 1960. After removing the tidal components, they found a general coastal drift of 0.10 knot (2.4 nautical miles/day) toward 267°T.

MILLER (1952) related the surface distribution of temperature and salinity and the routes of drift bottles released in May 1951 between Block Island and Cape Hatteras. His interpretation of the data, which developed detailed patterns of the temperature-salinity distribution and drift-bottle trajectories, suggested that the southerly coastal drift consisted of several distinct currents or eddies with branches and offshoots, merging or breaking away from the general drift. There was a tendency for the drift bottles to float along lines of similar salinities and temperatures. He noted that the influence of each estuarial contribution to the coastal circulation appears to have an appreciable effect on the current pattern, presumably due to the modification of the salinity-density distribution.

#### D. F. BUMPUS

Evidence of the escape of coastal water from the system was observed by FORD and MILLER (1952) and FORD, LONGARD and BANKS (1952) when they found a narrow discontinuous band of relatively cold and fresh water in the left margin of the Gulf Stream northeast of Cape Hatteras. They explained that the southward-flowing shelf water on the continental shelf turns eastward at Cape Hatteras and is entrained between the Gulf Stream and the slope water. FISHER (1972) has repeatedly observed entrainment of relatively cold low-salinity water by the Gulf Stream and subsequent formation of a cold filament adjacent to its northern edge northeastward of Cape Hatteras.

KETCHUM and KEEN (1955) applied techniques developed in the study of the flushing of estuaries to the Middle Atlantic Bight and reasoned that the accumulation of river water on the continental shelf affords a direct means of evaluating the rate of circulation and mixing when compared to the rate of discharge of the rivers. They summarized their findings as follows:

The depth mean salinities for the waters of the continental shelf between Cape Cod and Chesapeake Bay show a seasonal variation in the concentration of river water. The spring and winter accumulations are about the same, but about 25% more river water is present in the summer. The total volume of fresh water in spring and winter is equivalent to that produced by the rivers in about one and a half years. The extra accumulation in summer is equal to half a year's flow, and reflects, in part, the fact that the high spring flows of two successive years are present on the shelf at the same time.

There is a decrease in the average content of river water in the direction of the flow of the coastal current in spite of the addition of river water along its course. It is concluded that considerable transport of river water and of salt normal to the coast is necessary. The horizontal mixing coefficients normal to the coast are computed from the seasonal changes in salinity. They range from 0.58 to  $4.96 \times 10^6$  cm<sup>2</sup>/ sec, with the values for the decrease in salinity from spring to summer being smaller than those for the increase from summer to winter conditions. At both times, the values decrease with increasing depth and distance from shore.

In a further study, KETCHUM and CORWIN (1964) examined the pool of cold bottom water extending from south of Long Island to off Chesapeake Bay (BIGELOW, 1933) and found that the warming of the pool may result either from mixing with the warmer and fresher surface water or with the warmer and more saline slope water at a similar depth and density. They were also able to correlate the average salinity of the continental shelf in the area with the 6-month average flow of the Connecticut River. WHITCOMB (1970) found this cold pool as a continuing core of relatively cold water extending along the shelf edge from off Nantucket Shoals to the offing of Chesapeake Bay in September 1967. There was evidence of "calving" in which a large bubble of cold water had separated from the core and moved seaward as described by CRESSWELL (1967).

In their first experiment with a transponding buoy, BUMPUS et al. (1957) found that drift bottles set out south of Rhode Island in June move northward into coastal waters with an average drift of over 1 mile per day while the transponding buoy indicated an anticyclonic irrotational drift, erratic in its rates of movement and direction, suddenly speeding up and slowing down. This sort of drift may be repeated several times as the waters drift shoreward.

The effect on the temperature structure of onshore and offshore winds has had some notice from time to time. The maximum surface temperatures usually occur in August in the Middle Atlantic Bight, but the maximum bottom temperature is delayed to September, October or even November (BUMPUS, 1957b). WELLS and GRAY (1960) observed the depression of the surface water temperature along the northeast coast of North Carolina in July and August following southwesterly winds. They assumed an offshore movement of the surface water with subsequent replacement by upwelling of cold subsurface water. This event has been noted time and time again at the lightships along the east coast from Chesapeake to Ambrose (BUMPUS, 1957b; DAY, 1959a, 1959b, 1960a, 1963; CHASE, 1964, 1965, 1966, 1967, 1969a, 1969b, 1971a, 1971b, 1971c). The summer warming at the surface at one or several of the lightships appears to be interrupted from time to time. This is most conspicious during mid-July, August or September. This depression in the surface temperature results from the offshore movement of surface water and the intrusion along the bottom of cold saline water from offshore which influences the whole water column, i.e. upwelling. The intrusion follows a period of preponderant strong southerly winds. CHASE (1959) noted that these changes in the water temperature are the result of wind-induced advection rather than *in situ* modification of the water by the atmosphere. These changes are evidenced also by the rising and lowering of the summer thermocline as the warm surface layer becomes thinner or thicker. It was further pointed out by DAY (1960b)

Where waters of different characteristics are in juxtaposition, marked change in a given physical parameter at a given spot may be expected to occur with tidal periodicity. These abrupt changes are the result of advection rather than the modification of the water *in situ*. Prevailing winds can contribute to the dominance of one water type over others, at that location, for appreciable periods.

Northeast winds will drive Virginian coastal water past Cape Hatteras into the Carolinian province (see next section).

Numerous papers mention the uncertainty in applying the classical dynamic calculations to determine the current velocity in coastal waters due to the questionable validity of the three basic assumptions:

- 1. that there is a layer of no motion which may be used as a reference in computing the current,
- 2. that the observed distributions are in a steady state, and
- 3. that the boundary friction can be neglected.

Thus the dynamic topography developed from the temperature-salinity-depth determinations are seldom employed in shelf waters, or if so, most cautiously.

Direct comparison of the residual drifts and the density distributions in the Middle Atlantic Bight was carried out by Howe (1962) employing drogued buoys and later drogued transponding drift buoys (WALDEN, KETCHUM and FRANTZ, 1957) and the usual Nansen bottle sampling techniques, assuming that if the flow is geostrophic and velocity decreases with depth, the denser water will be on the left of the direction of motion and the lighter water on the right. He found during the early months of the year between Cape Cod and New York that the inshore waters within the 90-m contour were in a comparatively stagnant state, responding to short-term wind effects, whereas outside the 100-m contour, a permanent westerly drift persisted. His July-August cruise results revealed a southerly drift over much of the Middle Atlantic Bight culminating in a seaward exit between Chesapeake Bay and Cape Hatteras. His correlations between velocity and the change in  $\sigma$ , between stations were good. During November, when the waters were much less or not at all stratified, he found the drift reflected the local conditions, rather than the general nature of the flow. After comparing the summer with the late autumn data, he concluded that the changes in temperature throughout a section in summer play an important role in the resultant density structure whereas, in contrast, the salinity exerted an overwhelming influence in November.

The generalized report on the basis of all the drift-bottle data for 1948–1962 (BUMPUS and LAUZIER, 1965) stated:

#### D. F. BUMPUS

The westerly drift south of New England and Long Island and offshore off the Middle Atlantic States during the winter is chiefly inference. Inshore off the Middle Atlantic States the drift is southerly with a marked offshore component from Chesapeake Bay southward; an indraft from southeast and south of Nantucket Shoals develops during the spring toward southern Rhode Island. This surface drift does not extend into Long Island Sound. A westerly flow continues offshore south of Long Island, and the southerly flow off the Middle Atlantic States extends well across the shelf during the season, impinging on the outer banks of Cape Hatteras or turning offshore into the Gulf Stream along the edge of the continental shelf. Immediately south of the western two-thirds of Long Island there is an easterly counter-drift.

The indraft from western Georges Bank and south of Nantucket Shoals persists through June; from Great South Channel through August. The westerly and southwesterly drift toward Cape Hatteras narrows in August to the inner side of the shelf. It is not at all uncommon during the summer months to find the drift reversed off the Middle Atlantic States, especially next to the coast.

The indraft across Nantucket Shoals has disappeared by September, and the westward drift across Great South Channel has decreased considerably by the end of September. The westerly and southerly drift off the Middle Atlantic States is well-defined during the autumn close to shore, being interrupted by reversals less often than during the summer. As the autumn progresses, there are fewer returns from offshore, suggesting increases in the offshore component as the onset of winter approaches. Harrison *et al.* (1964) have recently demonstrated an anticyclonic eddy in the near shore area immediately south of Cape Henry, which our June, July, September, and October data suggest, but which we have not adequately demonstrated.

The exhaustive study by NORCROSS and STANLEY (1967) during 1963 and 1964 in the Chesapeake Bight showed that the surface circulation varies with the seasons dependent on the river effluents, the local winds, and the changes in the stability of the water column. While the prevailing surface-water drift was generally southerly, a northerly flow developed in summer when the water column was highly stratified and the southerly winds prevailed. Lack of returns from near the edge of the shelf indicated an offshore drift of the outer shelf waters south of 37°. Penetrations of the temperature-salinity barrier at Cape Hatteras permitted flow into the Carolinian province during times of northeast winds, but this breach is restricted to within a few miles of the coast.

HARRISON and PORE (1967) attempted to correlate the drift-bottle results of Norcross and Stanley with the local wind and runoff data. They concluded that the local winds may not always be a dominant driving mechanism; instead it might be useful to try to correlate the atmospheric pressure field over the western North Atlantic and to include the local density stratification in the prediction equations.

It was fortunate that the collection of temperature and salinity information at the lightships along the Atlantic coast (CHASE, 1969c) was set up in time to record the changes in salinity at these locations before and after the period of the low runoffs from the Delaware and Hudson River watersheds. As a result of the drought of the mid-1960s, reversals in the surface drift off the coasts of Virginia, Maryland, Delaware and New Jersey were observed in mid-summer with greater frequency than previously. These reversals were usually confined to a narrow belt close inshore. BUMPUS (1969) noted that 50% of the annual runoff from these watersheds normally occurs during March, April and May. During the drought the total annual runoff was reduced at the end of the third year to only 42% of average. The monthly mean surface salinities as observed at *Five Fathom Bank* Lightship were for the most part higher at the time of reversals than the average monthly means for the period of record. BUMPUS (1969) predicted

Surface current reversals in the Middle Atlantic Bight will not be expected between October and March. Although the salinity, hence the steric anomaly, gradient across the shelf will be small during this period, the dynamic gradient will be reinforced by the southerly component of the prevailing north-westerly winds. Surface current reversals may be expected at any time, April to September, when the wind is from the southern sector and the runoff during March, April, or May has been below normal, inasmuch as the salinity next to the coast will not be reduced sufficiently for the resulting pressure gradient to balance the drag of the southerly wind. Otherwise with normal or above normal runoff, the surface drift may be expected to remain southward.

A rough calculation indicates that a slope of 5 cm across the continental shelf off the coast of New Jersey is required to maintain a southward coastal drift of 5 nautical miles per day. When the slope is diminished, the southwesterly winds of summer are sufficient to slow down or even to reverse the drift.

The slope water current as observed at Site D ( $39^{\circ}20'N$ ,  $70^{\circ}00'W$ ; 50 km south of the shelf and 175 km north of the mean axis of the Gulf Stream; Fig. 1) has a net velocity to the west with the amplitude decreasing with depth. WEBSTER (1969) reported a persistent westward mean flow at all depths; the largest deviation from true west is  $11^{\circ}$ . The arithmetic mean of the vectors at the observed depths is within  $2^{\circ}$  of  $270^{\circ}T$  with speeds as follows:

10 m	13.0  cm/sec = 6  nm/day
100 m	7.0 cm/sec
500 m	3.8 cm/sec
l000 m	3.6 cm/sec
2000 m	1.6 cm/sec

The components of time-dependent velocities are not isotropic at all depths and time scales. The variance of the north component is greater than the variance of the east component at low frequencies (periods longer than a day) in the surface layers. Below 200 m the east component is larger. The dominance of the east-west component below 200 m may be due to an inhibiting effect of the nearby continental slope on the north-south component.

MILLARD (in prep.) has found a layer of bottom water, homogeneous in temperature and salinity, up to 28 m thick, at a number of stations at or near the shelf break south of Nantucket. A theory of shoaling internal waves (WUNSCH, 1968) suggests that internal wave kinetic energy is expected to concentrate along the bottom, the frequency concentrated being dependent in part on bottom slope. The intense shear thus developed near the bottom in a restricted location, near the shelf break, would cause the bottom water column to be mixed.

The use of sea-bed drifters (LEE, BUMPUS and LAUZIER, 1965) to study the residual drift along the bottom was commenced in U.S. waters in 1961. BUMPUS (1965) reported the results inferred from their returns for the area between Cape Cod and Cape Henlopen.

a. The residual drift is variable, not always in the same general direction, at times appears nearly reversed, but a general tendency is indicated in spite of these variations.

b. Offshore of a line drawn about 1/2 to 3/4 of the distance between the shore and the 50-fathom contour at depths of 30-35 fathoms, the tendency is toward an offshore drift. Inside of this line, the tendency is for the flow to be westerly or southerly with a component toward the coast. This line, whose location is not sharply defined and which may move on or offshore from time to time (probably seasonally) is a line of divergence.

c. There is a definite residual drift toward the mouths of estuaries. This is particularly evident south of Rhode Island where Narragansett Bay and Long Island Sound contact the sea.

d. The rate of the residual flow varies from 0.1 to 0.7 nautical mile per day, but more frequently 0.2 to 0.5 nautical mile per day (0.4–1.0 cm/sec).

NORCROSS and STANLEY'S (1967) sea-bed drifter results in the Chesapeake Bight are comparable. Although the direction of the bottom drift was variable, there was a pronounced trend toward the southwest to enter Chesapeake Bay at all seasons, but a northwesterly drift toward Chesapeake Bay from as far as 35 nautical miles southeast of Cape Henry at times of maximum fresh-water discharge. Few recoveries were obtained from drifters launched near the edge of the shelf, indicating an offshore bottom drift beyond 40 nautical miles from shore. Similarly an offshore drift was inferred from the paucity of recoveries of drifters launched east and northeast of Cape Hatteras. The speeds of drifters launched in autumn and winter were somewhat greater than those launched in spring and summer.

Corroborative evidence of the net drift toward and into estuaries is supplied by MEADE (1969). He reported

Bottom waters of the continental shelf move progressively into the mouths of estuaries, and they presumably carry bottom sediment with them. Beach sands move toward and into the mouths of some estuaries at rates of several thousand cubic meters per year. Distinctive mineral components in the lower reaches of estuaries suggest that the bottom sediments were derived from offshore.

## South Atlantic Bight

The study of the oceanography of the shelf waters of the South Atlantic Bight was relatively late in commencement. The Florida Current was of more obvious interest (PILLSBURY, 1891; BIGELOW, 1917; WÜST, 1924; STOMMEL, 1958). ISELIN (1936) diagrams three temperature-salinity sections across the area in May 1933. The first glimpse of the temperature-salinity distribution over the whole shelf is afforded by ANDERSON and GEHRINGER (1956–1960) and ANDERSON, MOORE and GORDY (1961). There had been some conjecture about the circulation on the part of navigators and geologists observing the movement of sand spits and capes, and biologists studying the alternations in the distribution of flora and fauna.

The first mention of counter currents inshore of the Florida Current is in a report of John White, who made a voyage from Florida to Virginia in 1590. In order to stay within the Gulf Stream, it was necessary to stand far out to sea because along the coast there was a countercurrent: "eddy currents setting to the south and southwest" (KOHL, 1868). Tuomey reported in 1848 that the Gulf Stream produced an eddy which washes the coast of South Carolina southward (ABBE, 1895). Abbe diagrammed a series of counterclockwise eddies fitting into the cuspate coastline of the South Atlantic Bight. RUDE (1922) did not find evidence of these eddies in currents measurements. Instead he found the surface drift responding to the local wind. PARR (1933) noted the winter-temperature barrier which occurs along the outer beach from Cape Hatteras northward, separating the Virginian and Carolinian faunal provinces. Indirect evidence that the barrier was frequently breached, permitting northern species to become established in Raleigh and Onslow Bays for a few months, was provided by WILLIAMS (1948, 1949), SUTCLIFFE (1950) and PEARSE and WILLIAMS (1951).

Seasonal events have also been noted near the southern part of this segment of the shelf. GREEN (1944) reported that the southerly winds prevailing in July and August cause an offshore transport of surface water in the vicinity of Daytona Beach, which is replaced by the colder subsurface waters of the Florida Current. The area thus cooled is extensive as indicated by the displacement of the air temperature–latitude curves for the months of July and August. By September the prevailing wind is northeast, and the temperature of both the surface water and the air returns to normal. The evidence of TAYLOR and STEWART (1959) indicates that this upwelling extends south of Cape Kennedy at least as far as Canova Beach. The lowered sea-water temperatures in July and August are accompanied by a depression in sea-level and southwest winds. Normal temperature and sea level are restored with the return of northeast winds.

HAIGHT (1942) reduced the tide pole data from nine locations to the hourly velocities, the harmonic constituents, the non-tidal currents by months and the wind currents. The non-tidal current appears strongest at Diamond Shoals, Cape Lookout and Frying Pan Shoals and least variable at Diamond Shoals. There is considerable variation in the residual current direction from month to month and for the same months of different years. The July and August current velocities are consistently the largest and are northeastward.

The first look at the temperature-salinity distribution over the North Carolina continental shelf was provided by PIERCE (1953) who found large-scale invasions of Florida Current water onto the shelf, interdigitated with shelf waters where large-scale mixing appeared to be taking place. BUMPUS (1955) reported that the southerly flowing coastal current, as is normally found in the Middle Atlantic Bight, is quite transient in the South Atlantic Bight; and when present, it is restricted to a very narrow band next to the coast. Instead a broad, very slow northerly drift is encountered over the shelf with frequent broad penetrations of Florida Current water onto the shelf, most deeply into Raleigh Bay, less so into Onslow Bay, and least into Long Bay. Six reasons were given for the lack of development of a dynamically driven coastal current.

- 1. No regular contribution of freshened water from north of Cape Hatteras.
- 2. River runoff per unit length of coast line (see Table 1) is less than half as great as that for the Middle Atlantic Bight, and the effluent becomes well mixed in the sounds and bays (ROELOFS and BUMPUS, 1963) before it exits to the shelf, reducing still further its effect on the steric anomaly.
- 3. The horizontal temperature gradient across the continental shelf during the winter months is more than sufficient to counteract the salinity effect on the pressure gradient.
- 4. The southwest wind blows parallel to the general trend of the coast, hence has a long fetch retarding or deflecting offshore a tendency for a south-flowing current.
- 5. The geography of the coastline with its deep cuspate shape and shallow offshore bars at the capes deflects currents moving in either direction offshore.
- 6. The adjacent encroaching Florida Current produces a frictional drag on much of the water over the shelf.

The actual breaching of the temperature-salinity barrier at Diamond Shoals was witnessed in January 1954, when Virginian coastal water penetrated well into Raleigh Bay (BUMPUS and PIERCE, 1955). Other occasions when this had occurred are noted by DAY (1959b, 1960a, 1963) and, as far as Frying Pan Shoals, by BUMPUS (1957b) and CHASE (1959). GRAY and CERAME-VIVAS (1963) found on two occasions (April and August 1962) a southwesterly coastal flow into Raleigh Bay. They suggest that a southerly flow is more prevalent in this region than had been suspected and is augmented by a persistent northeast wind.

The penetration of Florida Current water onto the shelf off North Carolina, either throughout the whole water column or as a wedge along the bottom (due to meanders and the subsequent drawing off of coastal water), are recorded by BUMPUS and PIERCE (1955). WEBSTER (1961), in his repeated sections through meanders off Onslow Bay during May and June 1960, discerned offshore running tongues of shelf water, similar in composition to that observed at Frying Pan Shoals 2 days earlier, being entrained in the shoreward edge of the Stream. The prominent components of the meanders have a period of the order of a week and amplitudes of 10 km. "Each meander resembles a sort of skewed wave motion and

consists of an intense offshore running current (time 1–4 days), followed by a broad confused flow onshore (time 4–7 days), then followed by another intense offshore current" (WEBSTER, 1961). BLANTON (1971) has measured compensatory shoreward bottom and offshore surface flows. He found the Gulf Stream water intruding across the shelf at the bottom a distance of 11 km in 25 hr (12 cm/sec). The surface  $24^{\circ}$  isotherm retreated offshore a distance of 15 km in 32 hr (13 cm/sec).

The surface current atlas of BUMPUS and LAUZIER (1965) is based on too little data to be very useful south of Cape Hatteras. With additional drift-bottle data it was discovered that there is a seasonal reversal in the drift south of Frying Pan Shoals. During the winter-spring the drift is northerly. In late summer and early autumn the drift reverses and flows southerly (BUMPUS and CHASE, 1964).

Studies of the residual drift off Cape Kennedy in March-April and August 1962 revealed a predominantly northerly drift during March and April of 12 cm/sec and a southerly drift during August of 4–5 cm/sec. The drift during the spring was, however, closely associated with wind direction. When the wind sprang up from the south, the drift quickly turned northward. When it hauled to the north, it took a large fraction of a day to reverse the current direction. The relation of the wind to current direction was less obvious in August (BUMPUS, 1964).

Finally, a very intensive study was carried out between the middle of Onslow Bay and Wimble Shoals, north of Cape Hatteras, during 1965–1967. STEFÁNSSON, ATKINSON and BUMPUS (1971) summarized the results as follows:

The wind-driven Virginia water transport past Cape Hatteras may markedly reduce the temperature and salinity, and accelerate the fresh water exchange. The runoff from the watershed north of Cape Hatteras, when driven south by northerly winds, especially during the peak period in early spring, has a much greater influence on the circulation in Raleigh and Onslow Bays than does that from the adjacent rivers south of Cape Hatteras. Meanders in the Gulf Stream may renew the shelf water with warm saline water. Intrusion of subsurface Caribbean water on to the outer part of the shelf takes place most frequently during the late summer following periods of southerly winds. It is postulated that during the cold part of the year, when the surface layers are only partly stratified, such an intrusion may lead to upwelling near the shelf break. Cascading from the shelf down the slope may occur during the cold season following periods of low air temperature.

NEWTON, PILKEY and BLANTON (1971) have graphically illustrated the summer and winter regimes of temperature and salinity in this area.

## CIRCULATION ON THE CONTINENTAL SHELF

## General monthly surface drift

During the years 1960–1970, 165,566 ballasted drift bottles were released in the waters of the continental shelf of the east coast of the United States. As of September 1971, 16,432 (9.9%) were recovered from the local shores of North America, with another 115 (0.1%) reported from overseas. This is a body of data comparable in size to that reported by BUMPUS and LAUZIER (1965) but with better distribution in space and time between South Channel and the Straits of Florida (see Appendix A).

The bottles used were 8-oz soda bottles, ballasted with a teaspoon of dry sand to assure they would float vertically with a minimum of free-board. A serially numbered, self-addressed red and white postcard requesting the finder to note date and location of finding was inserted in each bottle. The card promised a reward of 50 cents for the return of the requested information. The bottles were closed with cork stoppers. Five bottles were customarily dropped at each station except at lightships/lightstations where two bottles were dropped once a day.

All the drift-bottle data for both releases and those recorded were coded and the data placed on IBM punch cards per National Oceanographic Data Center format (ANONYMOUS, 1964). These keypunched data are on file at the Woods Hole Oceanographic Institution and the National Oceanographic Data Center. The data were arranged by month of release, irrespective of year, and by 10 min of latitude and longitude of release. For those recovered, the data were further ordered by the elapsed time between release and recovery.

The number recovered on North American shores from each 10-min rectangle were enumerated and compared to the number released. The percent return was smoothed by grouping these numbers from 10-min rectangles into areas of 30 min. Percent recovery contours were drawn on charts for each month.

The mean direction of drift from each 10-min rectangle each month was determined by averaging the azimuths between launch position and recovery site. Where the data suggested there was more than one preponderant direction of drift from a rectangle, occasioned by the drift being different during a given month of one year from the same month another year, or different within a given month, more than one azimuth of drift was recorded. These directional arrows were added to the monthly charts. The speeds of the fastest drifts were recorded, and the lengths of the arrows were drawn as a function of these fastest drifts. Thus, for those drift bottles which had lain on the shore for a long time before being found, the elapsed time between release and recovery was ignored.

Inasmuch as we know only the location of release, the site of recovery, and the elapsed time between release and recovery, a further step was taken to infer the direction of drift, using the field of data and working upstream. Velocity vectors were plotted on the final charts for those bottles which travelled the shortest distance. These vectors influenced the route of the next succeeding parts of the field and so on to those which travelled the longest distance. This step appeared to produce a fairly reasonable monthly plot of the field of motion.

Figures 3–8 of the inferred monthly surface drift are drawn with the following notation:

- 1. A dot ( $\cdot$ ) in the center of each 10-min rectangle from which no bottles were recovered.
- 2. A star (\*) in the center of each 10-min rectangle from which bottles drifted to an overseas location.
- 3. An arrow (→) indicates the inferred direction of drift from each rectangle; the length of the arrow is a function of the fastest drifts recovered.
- 4. Recovery rates of 1, 10, 20, 30, 40 and 50% are contoured; those greater than 50% have not been drawn.

The charts of inferred monthly surface drift added very little to our previous understanding of the circulation in the Gulf of Maine when compared to that presented by BUMPUS and LAUZIER (1965). This results from the diminished effort in this area, particularly over Georges Bank during this period. One can discern the development of the two-gyre system; the Maine counterclockwise eddy and the Georges Bank clockwise eddy are reaffirmed, but the data are too limited for a satisfactory development of the seasonal progression.

With a denser distribution of data in the Middle Atlantic Bight, some additional interpretations are now available. During December–February (Figs. 8 and 3) the recovery rate of drift bottles is exceedingly small (<10% adjacent to the shore) so there is a minimum of













direct evidence on surface velocities. The few velocities plotted reveal a general southwesterly drift with speeds approaching 12 nautical miles per day. Off the mouths of estuaries the drifts are not at all consistent: flowing with the main stream on occasion, entering the estuaries on others, or reversing and flowing easterly, as off New York Harbor. We can guess that the drifts over the remainder of the shelf are nearly parallel with those figured but somewhat slower because all the plots of the density distribution we have seen for these months show the surface  $\sigma_t$  isopleths more or less parallel with the coastline, with the gradient weakening offshore. Strong, persistent winds can markedly modify the geostrophic flow during this period.

The drift bottle recovery rate improves considerably over much of the shelf during the remainder of the year except south of the eastern end of Long Island. It is possible to infer from this a two-cell system in the Middle Atlantic Bight (e.g. Fig. 5). The eastern cell received indrafts from offshore south of Nantucket Shoals and from Georges Bank across South Channel during the spring months, which converge south of Massachusetts and Rhode Island with a compensatory offshore flow east of Hudson Canyon. The position of this offshore flow, as evidenced by the less than 1% recovery zone, shifts from time to time. Its area appears largest and closest to shore when the indraft south of New England is most vigorous. Another cell develops between Hudson Canyon and Cape Hatteras, although indrafts from offshore have not been as frequently observed as those south of Nantucket. This cell receives the outflow from the major rivers emptying into the Middle Atlantic Bight, including the Connecticut, and continues with a southerly flow toward Cape Hatteras occasionally reaching 15 nautical miles/day, but for the most part restricted to 10 nautical miles/day or less. This southerly flow is interrupted during the summer months by slow northerly wind drifts, less than 5 nautical miles/day as noted earlier (BUMPUS, 1969). The maximum distance offshore where reversals have been noted is 40 nautical miles. Usually they are confined to and are strongest within 30 nautical miles of the coast. During the autumn the indraft south of New England weakens, and the zone of no recovery recedes from near the 100-fathom line and converges toward the coast, where the southerly flow prevails except for a few reversals within 5 nautical miles of the shore.

An offshore flow of shelf water between the offing of Chesapeake Bay and Cape Hatteras is not evident in these data, other than on the occasions when the 1% recovery contour is proximate to the coast during January, February, August, October, November and December. There is evidence, however, of a drift close to the shore past Cape Hatteras into Raleigh Bay during March through June (Figs. 4–5) and possibly in July and September (Figs. 6–7). It is conceivable that the drift, which impinges on the shore as it flows southward during the warm half of the year (e.g. Figs. 5–7), has a component slightly to the left during the cold half, when the northwest wind prevails, missing contact with the shore; hence, it either flows into Raleigh Bay or turns offshore at the southern end of the Middle Atlantic Bight. There is probably considerable exchange of shelf water and slope water along the outer edge of the shelf at all seasons. The surface isotherm charts issued monthly by the Coast Guard Airborne Radiation Thermometer Program show these large-scale exchanges taking place.

The recovery rate of drift bottles from the South Atlantic Bight is poor during the colder part of the year (Figs. 3 and 8), as it is over other portions of the shelf. The data suggest a northerly component to the drift from off Georgia to Cape Hatteras in January (Fig. 3), with a southerly flow along the Florida coast to where the shelf narrows south of Cape Kennedy. The highest rate of return during January is from the vicinity of Cape Kennedy. The drifts are similar in February (Fig. 3) with the exception of a weak (not more than 4 nautical miles/day) southerly drift close to the shore in Raleigh and Onslow Bays. This southerly drift broadens in March (Fig. 4), is augmented by Virginian coastal water, and may flow close to the shore all the way to northern Florida. The drift is northerly from the Straits of Florida over all the shelf as far as Cape Kennedy and is northerly over the outer part of the shelf to Cape Lookout. These drifts are on the order of 5 nautical miles/day. The intrusion of Virginian coastal water in April penetrates only as far as Raleigh Bay. The southerly drifts over the innerpart of the shelf are discontinuous as a northerly drift prevails over most of the shelf from Cape Kennedy northward. In the offing of Cape Kennedy the drift turns southerly.

The weak northerly drift which prevails over much of the shelf in April (Fig. 4) gives way in May (Fig. 5) to a stronger (as much as 7 nautical miles/day) southerly drift from Frying Pan Shoals southward. This is augmented by indrafts from the Florida Current all along the 100-fathom contour. From Frying Pan Shoals northward the drift has an onshore component which is deflected northerly at one time, southerly at another.

By June (Fig. 5), with the recovery rate reduced, the northerly drift off Florida has resumed over the outer part of the shelf to Frying Pan Shoals, with a southerly drift continuing inshore off Georgia. Virginian coastal water flows into Raleigh Bay past Cape Hatteras, augmented by an indraft of Gulf Stream water.

With the recovery rate still modest in July (Fig. 6), the drift off North Carolina is now northeasterly. The Virginian coastal water indraft has been pinched off. The drift off Cape Kennedy is still northerly, but a southerly drift begins over the mid-part of the shelf off Georgia. This southerly drift expands during August and continues during September (Fig. 7), involving the whole shelf from Frying Pan Shoals southward. Florida Current water becomes entrained in the shelf drift, which commonly reaches speeds of 9 nautical miles/day. Off North Carolina the direction of drift is intermittent, northerly at times, southerly on other occasions. By October and continuing into November and December (Fig. 8), the southerly drift is restricted to off Georgia and Florida; the remainder of the shelf exhibits a very poor recovery rate, insufficient to determine the direction and velocity of drift.

It would appear that two conflicting systems are at play here, a geostrophic current interrupted by invasions of the Florida Current. The geostrophic current tends to flow southerly and does so successfully in May, during late summer, and early autumn from Frying Pan Shoals southward. It is interrrupted frequently by invasions of the Florida Current riding up over the shelf carrying the surface water northward. On those occasions when the recovery rate is poor in the South Atlantic Bight, one can generally assume the surface water has been forced out by a meander of the Florida Current, and the drift bottles have been carried north past Cape Hatteras and out of the immediate system. It would appear that the meanders are more frequent and more successful in accomplishing this total exchange of water north of Frying Pan Shoals.

## General monthly bottom drift

During the years 1961–1970, 75,485 sea-bed drifters (LEE, BUMPUS and LAUZIER, 1965) were released in the waters of the continental shelf of the east coast of the United States. As of September 1971, 12,008 (15.9%) were recovered from the shores of North America, 1102 (1.5%) of which were found in estuaries. Others totalling 2106 (2.8%) were

recovered from the bottom by fishermen (see Appendix B). Except for the discussions of the bottom drift off southern Nova Scotia (LAUZIER, 1967), off the coast of Maine (GRAHAM, 1970), in the Middle Atlantic Bight (BUMPUS, 1965) and in Chesapeake Bight (NORCROSS and STANLEY, 1967) a description of the bottom circulation has not been reported before.

For the most part the sea-bed drifters were released in bundles of five at the same time and place as the drift bottles reported above, though none were released regularly at the lightships/lightstations. The bottom drifter data were coded, keypunched, arranged, and treated in the same manner as the drift bottle data. Figures 9–14 of the inferred monthly bottom drift are drawn with the following notation:

- 1. A dot (·) at the center of each 10-minute rectangle from which no sea-bed drifters were recovered,
- 2. An arrow  $(\rightarrow)$  indicates the inferred mean direction of drift from each 10-min rectangle from which drifters were recovered; in these charts speed is not indicated, and the arrows are all the same length,
- 3. Recovery rates of 10, 20, 30, 40, 50, and 60% are contoured; recoveries greater than 60% have not been contoured.

It may be unrealistic to draw monthly charts of bottom drift for the whole continental shelf inasmuch as the speeds of drift are on the order of tenths of a mile per day and it may take a drifter from the mid-part of the shelf one-half year to a year or more to strand. However, those launched inshore are recovered in shorter times, and it does seem worthwhile to reduce the data to charts on a monthly basis to compare with the surface drifts, chiefly to see if there is any discernible seasonal cycle in the recovery rate.

A persistent and continuous bottom drift of  $0.5\pm0.2$  nautical mile/day extends toward the southern tip of Nova Scotia and the eastern side of the Bay of Fundy from Browns and La Have Banks east of the Northeast Channel. This is in agreement with LAUZIER'S (1967) findings. Along the western side of the Gulf of Maine the drifts next to the coast tend to flow directly ashore, whereas farther offshore the drift is more nearly parallel with the coast in a westerly direction. Fewer than 10% of the drifters are recovered from the deeper parts of the Gulf of Maine and from Georges Bank. The drifts in the deep parts (>100 m) are less than 0.1 nautical mile/day whereas those for Georges Bank are on the order of 1 nautical mile/day. A line of divergence occurs at Northeast Channel with northerly drifts north and east of the channel and westerly drifts south of it. In general, the drifts over Georges Bank follow a clockwise rotation around the shoals with a net drift to the west and across Great South Channel.

The recovery area expands to approximately the 100-fathom isobath off southern New England, to include on the average the inner half of the Middle Atlantic Bight shelf within the 20% contour with the 10% contour lying slightly outside of it. From Great South Channel to south of Rhode Island the drift is northwestward  $0.7\pm0.2$  nautical mile/day. South of Long Island the drift inshore tends northwest to northeastward  $0.7\pm0.2$  nautical mile/day. A high return rate (36% annual average) is encountered in the New York Bight, with drifts directed toward Long Island and New Jersey shores including 3% of the recoveries from estuaries mostly from lower New York Bay.

The bottom drift converges on the coast of New Jersey from  $38^{\circ}30'$ N northward to approximately 40°N. From  $38^{\circ}30'$ N southward the bottom drift turns left and tends at  $1.0\pm0.5$  nautical mile/day toward the mouth of Chesapeake Bay and the outer banks of













F 145

North Carolina where recovery rates are high. In contrast to the interpretation of NORCROSS and STANLEY (1967) who suggest that sea-bed drifters launched east and north of Cape Hatteras are seldom recovered, it would appear that the recovery rate over the outer part of the shelf from 40°N toward Hatteras is a function of the distance from the edge of the shelf. In general, the 10% and 20% recovery lines parallel the 100-fathom contour. The data support BUMPUS's (1965) inference of a tendency for an offshore bottom drift over the outer part of the shelf with an onshore component over the inner one-half to two-thirds of the shelf. The 10% and 20% recovery contours oscillate over a fairly wide band (approximately 40 nautical miles) but not with any discernible seasonal progression. Lowest recovery rates occur and recovery contours are closest to the beach during the first half of the year, while highest recovery rates prevail and the contours are farthest from the beach during the second half of the year.

The bottom drift over the shelf of the South Atlantic Bight appears to divide into three general regimes: the offing of Raleigh, Onslow, and Long Bays; the offing of southern South Carolina and Georgia; and the offing of Florida. Off the Carolina Bays the mean bottom drift is northerly 0.7+0.4 nautical mile/day over the outer and mid-parts of the shelf, with a tendency to flow directly toward the coast or southerly inshore. There is a suggestion of convergence toward the capes, especially Hatteras, Lookout and Fear. Off southern South Carolina and Georgia the direction of drift is much less constant, tending northerly during some months and southwesterly  $0.8 \pm 0.2$  nautical mile/day during others, but without any apparent consistent pattern. These drifts off the Carolinas and Georgia may reflect the effect of the bottom indrafts of the Florida Current over the shelf, which seem to occur with greater frequency northeast of Frying Pan Shoals, but which may also occur with less frequency along the remainder of the shelf. Off Florida the drift is northerly: 1 nautical mile/day over the outer shelf and slope with a confused, sometimes north, sometimes south, drift between Cape Kennedy and Jacksonville at  $0.9 \pm 0.2$  nautical mile/day. There is a definite convergence toward Cape Kennedy from offshore and frequently a southerly drift inshore south of the cape.

Moderate recovery rates may be encountered in the Carolina Bays. In general, poor recovery rates prevail off the sea island coast with high recovery rates off Cape Kennedy and southward, where upwelling has previously been mentioned. It appears significant that the bottom drift is weak and inconsistent near and toward the coast which has no barrier beach, is more constant where the barrier beach is well formed, and is well developed with a high return rate where the capes appear in equilibrium. These drifts, with their sediment load, may have some bearing over long periods of time on the development of barrier beaches and capes.

## DISCUSSION

The recovery of sea-bed drifters along the shores of the east coast has shown no significant annual cycle. Sea-bed drifter recoveries during the winter months are nearly as numerous as during the summer, which is in contrast to the recovery of drift bottles. Until sea-bed drifters were broadcast in shelf waters on a year around basis, some people inferred that few bottles were returned during the winter months because the beaches were seldom frequented during that period. The frequency with which sea-bed drifters are recovered and returned during all seasons of the year leads one to believe that the shores are well patrolled all year long. This is certainly so along the sandy stretches where surf fishing and beachcombing provide active recreation. The rocky coasts, less easily walked on, are patrolled less well. Bottles are easily broken on these shores, and sea-bed drifters lodge in crevices. The marshy coast of the Georgia sea islands may not be well patrolled either, as it is far from resident populations and difficult to get to. But the fact that sea-bed drifters are recovered continuously during the year and drift bottles chiefly during the warm half of the year is evidence that the drifts reflect the thermohaline circulation modified by the wind. Where the salt wedge is well developed over the shelf, as in the Middle Atlantic Bight and off North Carolina (see Figs. 2 B and C and BUMPUS, 1965), the shoreward component of the bottom drift extends farthest from the shore. Where the salt wedge is less well developed, as off Georgia, the shoreward bottom drift is restricted closer to the coast. The surface drift tends to flow nearly parallel to the shore, having a slight shoreward component during the warm half of the year and a slight offshore component during the cold half when the winds are more consistently from the west and when the horizontal density gradients are weakest.

The validity of the percent recovery contours is questionable: we know they represent a minimum. What percent of bottles and drifters are found but not reported? Some become decorations in the bars of seaport towns. One bottle was acquired at an auction of a deceased person's effects in Tennessee. It was probably picked up on the coast of Maine where its mates were found. Some bottles may strand and become buried by wind-blown sand. Some of Henry Bigelow's bottles launched in New York Bight were recovered within 30 days near Cape Hatteras. Thirty years later another of the group was found in the same place. It had been buried and later uncovered by a winter storm. Some enter inlets and strand in marshy areas which are seldom frequented except during water-fowl shooting season. As of July 1972 very few bottles have been recovered since the "books were closed" on the statistical analysis in September 1971. Considerably more sea-bed drifters, several hundred, but only a fraction of a percent of the total released, have been recovered during the same interim, reflecting the slower speed of the bottom drift. Altogether these returns would probably not add more than a few percent to the overall recovery rate.

More significant than the actual percentages of recovery is the areal distribution of recoveries. The large expanses of no recoveries on the charts of surface drift compared to the same areas where the toward-shore bottom drifts are indicated show the difference in the surface and bottom circulation. The concentration of contours in certain places on the charts with sparse recovery in others indicates the areas where there must be a consistent shoreward component. Certainly we can have less dependence on the indicated directions of drift in the areas of less than 10 or 20% recovery than on those of 30, 40, or 50% recovery.

The drift-bottle and sea-bed drifter program was useful, but now a better understanding of the causes of these flows is needed. A discussion of possible ways to go about that follows.

## HOW SHOULD WE PROCEED ON THE CIRCULATION PROBLEM?

So far, not enough has been accomplished to describe the circulation on the continental shelf. We have only a general estimate of the flushing time of one portion of the shelf (the Middle Atlantic Bight, KETCHUM and KEEN, 1955). We have a minimum of information on the direction of drift during the winter months. We have no idea of the frequency of frontal disturbances in the vicinity of the 100-fathom line, except off Onslow Bay, nor the rate of exchange of shelf and slope water along this front. We do not know what role canyons play in the exchange of salt and fresh water across the shelf. It would seem that information of

this sort is now necessary as man seeks an easy and workable solution to his problem of waste disposal, including heat. Conservation-minded citizens caution against the use of the sea for disposal purposes even when the materials they wish to dispose of are the same as those which naturally occur there, i.e. biological wastes. They do not have sufficient scientific evidence to support their views. They extrapolate from the eutrophication of Lake Erie and other lakes. Do we have enough evidence to assess properly the rate at which nitrate, phosphate, carbon, etc., can be added to the ocean without causing eutrophication, or creating other disturbances? What effect will the heated effluents of power plants have on the circulation as well as on the life histories of the native flora and fauna? The present Governor of the Commonwealth of Massachusetts is against prospecting for oil on Georges Bank and elsewhere on the continental shelf until he is absolutely certain that an oil spill will not despoil the beaches and marshes. Can we put his fears at rest and begin to find and exploit this valuable resource? A better understanding of the circulatory exchanges is required to assess these problems effectively.

ISELIN (1955) asked: "Why have we at Woods Hole seemingly so long avoided problems in coastal circulation?" He pointed out then that more vigorous uses of the classical methods of oceanography, measuring the temperature and salinity distribution, would not solve the circulation problem. He suggested the use of continuous observations at well-selected points across the shelf. CARTER (1969) agreed with Iselin. He recommended

A program of collection of oceanographic and meteorological data, such as temperature, salinity, wind, water speed and direction, on the continental shelf of the east coast of the United States which will provide time histories of these variables at specific points or sections. The observations should be made over a long enough period of time to reveal all periodicities up to and including the annual. Although such a program could be carried out by multi-ship operations, moored arrays of instruments capable of sampling the entire water column would probably be better. Such a program should permit evaluation of wind, river inflow, tide, and internal waves as transport mechanisms.

Part of the reason we have been so slow in response has been the lack of suitable in situ instruments to collect the required data. ISELIN (1955) mentioned the Instrument Recovery Buoy. It took 10 years for that concept to evolve into a mooring release for submerged current-meter arrays. ISELIN (1955) mentioned that improved navigation was then available. Loran coverage of most of the continental shelf is satisfactory. For some types of position fixing, the accuracy of Loran is not sufficient. VLF Relative Navigation (STANBROUGH and KEILY, 1964) is now available as are Hastings-Raydist, Decca Navigation System, Hi-Fix, and others. Recording thermographs of various types with the required accuracy are available. Recording current meters of the Savonius rotor type, available for 10 years (RICHARD-SON, STIMSON and WILKINS, 1963), have been much improved, and the analytical software is now excellent. It would seem that we are ready to begin to heed Iselin's advice. The Ad hoc PANEL'S (1971) advisory report to the Coast Guard (regarding the Coast Guard's role in the Coastal Zone Baselines and Monitoring for Pollution and Environmental Quality Program) recommended among other things that each Coast Guard District be equipped with sets of moored current meters, which they should deploy quarterly for a minimum of 10 days along selected sections extending outward across the continental shelf. (The 10-day minimum was a reflection of the problems with mooring hardware. Hopefully, a minimum of 29 days, one lunar cycle, could now be invoked.) This program is to be implemented by the Coast Guard Oceanographic Unit soon.

What options are available for solving the circulation problem (determining net current velocities and transport, exchanges of coastal and slope waters) on the continental shelf

during the next few years? A combination of approaches is needed: a continuum of measurements at strategic locations, chiefly across fronts, with active field programs in the vicinity of or between the continuum stations (sections) in order to understand the field within which the *in situ* continuum measurements were located.

## I. Continuum measurements

The temperature and salinity measurements at lightships/lightstations, which commenced in 1955 and are now being conducted by the Coast Guard Oceanographic Unit, are an example of continuum measurements. They are made once a day with the exception of bottom salinities which are sampled twice per week. Although the lightships/lightstations are not necessarily located for scientific purposes (they are deployed as aids to navigation), a considerable amount of worthwhile information has been gathered, used, and reported during their 16 years of operation as "oceanographic observation posts". This program should be continued, and the sampling procedures upgraded with continuous sensing units, including current measurements, at a number of depths.

Instruments can be planted as part of a moored array with a surface float or submerged buoy recoverable by acoustic triggering of an anchor release. Instruments may be planted by skindivers in depths to 12 fathoms, allowing for a few minutes' work on the bottom, with no decompression required. (The thought here is: if the mooring blocks of certain aid-tonavigation buoys were prepared with a hole in them suitable for containing a recording thermograph, skindivers could insert and recover these instruments on a monthly basis. Similar equipment could be attached to a rack on the side of the buoy. Thus, surface and bottom temperature records would be acquired.) Beyond 12 fathoms, ships or mini-subs can plant and recover equipment. Acoustic pingers (MARQUET, WEBB and FAIRHURST, 1968) or bottom sonar reflectors are available for submarines to return to the instrument site.

(a) Temperature and salinity measurements. Thermographs are presently available off the shelf. Battery-operated temperature-conductivity-depth recorders, modified to record time rather than depth, would provide two parameters rather than one. The sensitivity of the conductivity cell is frequently reduced by fouling, hence cleaning the electrode with some regularity, possibly weekly, would be required. In the background narrative section mention was made of transient or periodic excursions of Gulf Stream, slope, coastal, Virginian, or upwelling water. Recorders planted in pairs, at the bottom and surface on existing navigation aids, or several arrays of recorders moored at intervals along a section or sections where fronts and excursions are known to exist would provide information on their frequency.

(b) Current measurements. The Richardson current meter has engendered a great flurry of current measurement in the deep ocean. It has been used successfully to replace the Roberts radio current meter in tidal surveys in harbors and estuaries. The recommendation to have the Coast Guard carry out a continuing program of current measurements along sections in cooperation with the research community is a good one and should be encouraged.

Is the Richardson current meter the ultimate current meter? Most probably not. A current meter with no moving parts is an interesting concept: the body of the current meter would orient with the direction of the flow relative to an internal magnetic compass, and the deflection of the supporting structure (vertically mounted wand or shape) would convey through a pressure transducer the speed of the current. Possibly the two-component electro-

magnetic flow meter described by OLSON (1972) could provide the speed input. HANFF (1971) appears to have anticipated these requirements.

The measurement of electrical potential to determine flow speed (von ARX, 1950) has been used successfully across the Straits of Florida (WERTHEIM, 1954), in the passages of Passamaquoddy Bay (TRITES and MACGREGOR, 1962), across Woods Hole Harbor (MANGELSDORF, 1962), and has been sensing the tidal current oscillation in Vineyard Sound for nearly a year. Certainly this method can be used to measure volume transport in and out of the estuaries by running a conducting cable between the capes guarding the mouth of the estuary. Corollary measurements will be necessary if the flow speed is not uniform with depth. Sanford (DREVER and SANFORD, 1970) is developing a free-fall electromagnetic current meter for use as a current profiler. This instrument clearly records small-scale velocity variations as a function of depth. Is it possible to develop a bottom-mounted electromagnetic current meter which can measure the volume transport above a unit area and deploy several of these along a section across the shelf and over the edge of the slope? Or can a mile or so of wire be stretched across the sea bottom in an "X" pattern with electrodes at each end and a recorder at the center, which can measure the volume transport above that square mile?

We cannot as yet adequately measure the slope of the sea surface. Even with an order of magnitude improvement over present capabilities (von ARX, 1969; von ARX, DEAN and SPENCER, 1971), a precise leveling method at sea or satellite altimeter could not measure the very slight slope  $(5 \times 10^{-7})$  of the sea surface across the continental shelf.

The slope across the shelf must be continually changing due to changes in wind, barometric pressure, tide, current and river runoff. One imagines the slope may be greatest and fluctuates most next to the coast, that it is more nearly level across the center of the shelf and quite large again near the 100-fathom line. Bottom-mounted tide gauges would not measure the slope of the sea surface, but pairs of tide gauges would reveal, after suitable filtering and integration, changes in the slope. With corollary measurements one could learn what these changes in slope signify. HICKS, GOODHEART and ISELEY (1965) have used a bottommounted, pressure-induced tide gauge on the shelf south of New England. This type of gauge can measure changes of tidal height to 0.1 ft. COLLAR and SPENCER (1971) describe another gauge accurate to  $\pm 1$  cm but with drift problems. SNODGRASS (1968) has developed a much more sensitive instrument for use in the deep ocean capable of measuring changes in height of 1 mm. Other highly accurate gauges are in the process of development or field testing (BAKER, 1971).

It may be that various effects altering sea level at a given location are so complicated as to make it impractical to unravel and discern those effects due to current. We will not know until we try. One lightstation, Savannah, is within sight of land. A tide gauge could be mounted on that structure, another nearby on the bottom, and a third at the shore; and the three could be compared. A reversal in the current drift occurs in this region from time to time; hence the tide gauge ashore should indicate a rise relative to those offshore when the drift is southward.

(c) Instrument accuracy. The study by BRANHAM and COSTE (1971) on the availability of instruments for the Coastal Zone Baselines Monitoring Program indicates that there are instruments which can measure the desired parameters but not always at the best accuracy. While the accuracies quoted in the brochures can be obtained under laboratory conditions, the actual accuracies are degraded with time in the field environment. It seems to this writer that we can accept, at this stage, *in situ* measurements of lesser caliber than required for active measurements. One can trade off the ultimate in accuracy for a time sequence which will provide some insight into the general range and frequency of time dependent activity.

## II. Corollary measurements

The continuum measurements mentioned above can be enhanced by, or in some cases require an active program of, other measurements by providing data which can be related to the continuum. It is still necessary to make temperature and salinity measurements in sections across the shelf from time to time in order to assess the fluxes of fresh water and salt. The recently developed techniques of PIERCE, NELSON and COLQUHON (1971) and MEADE, SACHS, MANHEIM, SPENCER and HATHAWAY (in prep.), in which water types are identified from their suspended mineral content, should be extended. This could provide a method, coupled with the T/S relation, of identifying river water, shelf water and slope and Gulf Stream waters and their mixtures, affording a means of examining the exchanges of these water types.

Current measurements. It would appear that we have learned just about all we can with drift bottles on this continental shelf. Certainly we gain a minimum of information during the winter months when we recover only 1% of the drift bottles whereas 21% may be recovered from August releases with an annual average of 10% for the whole shelf. The seabed drifters provide a higher percentage return (19%) being much more consistent in their monthly rate of return. Even so, drift bottles and sea-bed drifters provide only "a birth notice and an obituary with no biography". We really need to know much more detail about the tracks of a field of water followers inter se and in toto to truly exploit the Lagrangian method. The experiments of Howe (1962) with radio-responding drogued buoys produced rewarding results. However, the research vessel had to expend so much time tracking the buoys that it was not possible to field as many buoys as needed, nor could fixes be made at regular intervals. We have had one experience with a drifting IRLS buoy (EWING and STRIFFLER, 1970), but have abandoned further development because of the high cost, limited number of buoys which can be interrogated by a satellite in one area, number of fixes per buoy being limited to two per day, and lack of good communications. The latter weakness could be improved, but the others still remain. One other IRLS buoy has drifted for 21 days in September 1970 southwesterly at 4.2 nautical miles/day in the slope water off the Delaware Capes (CRUMPLER and BIVINS, 1972).

F. Striffler of the W.H.O.I. Engineering Department is in the process of developing a much less expensive "free-drifting buoy location system". This system will use the VLF navigation principle, in that each buoy will carry an ultra-stable oscillator, as will each receiving station. By comparing the phase of the buoy signal to that of the receiving station, the range from the station to the buoy may be determined. At least two shore stations are required. Ten to twenty buoys equipped with drogues at several set depths could be located every 2 hr for a period of 2 months, with a position error at the end of that time of about 10 miles. A follow-up program is envisioned to configure the buoy and drogue shapes so as to have the best grip on the water, and be least affected by wind. If Howe's correlation between the direction of drift and the density distribution can be further substantiated, and the speed related to the density gradient, we will be making real progress. Can we ascertain at what stages the geostrophic drift gives way to wind influence?

When continuum indicators are employed, such as bottom-mounted tide gauges, some means must be employed to determine the nature of the current profile, i.e. with a winch-lowered current meter over a tidal cycle. Two other methods are presently available. The "transport floats" of RICHARDSON and SCHMITZ (1965) have been used successfully in the Straits of Florida. It is possible that the shallow depth of the shelf and slope and the slower currents, reducing the distance and time from drop to recovery, will make the navigation limiting to this approach. The free-fall GEK (DREVER and SANFORD, 1970; SANFORD and DREVER, in prep.) appears to hold more promise.

#### **III.** Preliminary experiments

Quite a suite of tools and methods have been briefly outlined above. There are undoubtedly problems with each one which have not been mentioned or even visualized at this writing. Hence a test range or section is suggested. Research submarine *Alvin* has presently established a test site on the continental slope in the vicinity of Block Canyon (W.H.O.I. Permanent Bottom Station) for some biological experiments. It is suggested that the site be extended to a test section from Block Canyon to Montauk Point. Various continuum measurements would be made at selected points along the section, and corollary techniques would be employed between these points in order to relate the time dependent parameters. As experience and understanding progress, the successful techniques can be conducted over other parts of the continental shelf and slope.

There are problems which must be considered which do not prevail in deep water off the shelf. The east and west Nantucket/Ambrose Sea Lanes, with a submarine submerged transit lane in between, occupy a band 13 miles wide. These lanes will need to be avoided. The rest of the shelf area out to about 100 fathoms is frequented by fishermen of several nations. The otter trawls sweep the bottom night and day the year around where there is a possibility of a profitable haul. Strings of lobster pots stretched across the bottom provide another hazard. Unlighted buoys marking the site of instruments on the bottom are unintentionally run down or their mooring lines cut by trawl warps at night. Thus instruments set out on the bottom cannot survive long if positioned protuberant from the bottom. They must be so shaped that they lie flat or nearly so (turtle shaped), or dug into the bottom so that fishing gear will not tangle in them or sweep them into the trawl. An alternate possibility lies in the use of lighted buoys, similar to aid-to-navigation buoys, marking the site where instruments are deployed in close proximity. The surface buoys could be instrumented with temperature and weather sensors to enhance the input of needed information.

### IV. Auxiliary input

Weathership Hotel (38°00'N, 71°00'W) and XERB buoy (36°31'N, 73°31'W) are providing extensive weather information from the slope water region. The weather radars at Chatham and Wallops Island provide good coverage of rainfall over the area. Streamflow computations as provided by BUE (1970) will be required for freshwater flushing studies. The Coast Guard Oceanographic Unit's monthly Airborne Radiation Thermometer flights provide useful synoptic surface temperature data. They should be continued with greater frequency. It would seem that a continuum and corollary field program should be the next procedure to further our knowledge of the circulation and fluxes across the continental shelf.

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