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On Estuarine and Continental-Shelf Circulation in the Middle Atlantic Bight

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7.1 Introduction

We shall attempt in this chapter to trace the development of ideas about circulation over the continental shelf in the Middle Atlantic Bight and in the major estuaries which drain into the Middle Atlantic Bight. The term *Middle Atlantic Bight* refers to the curved section of the continental shelf off the eastern United States stretching between Cape Hatteras to the south and Cape Cod and Nantucket Shoals to the northeast. The New York Bight is a subsection of the Middle Atlantic Bight and refers to the shelf region stretching between the New Jersey and Long Island coasts. A schematic version of Uchupi's (1965) topographic map is shown in figure 7.1, indicating both the general shape of this shelf region plus the names and locations of the major estuaries and key positions discussed in the text.

We have decided to focus this review of estuarine and shelf cirulation on the Middle Atlantic Bight and its estuaries for several reasons. The major Middle Atlantic Bight estuaries have been extensively examined and have provided several important case studies in the development of new ideas about circulation and turbulent-mixing processes in moderately stratified coastal-plain-type estuaries. These estuaries and the adjacent continental shelf border on one of the world's largest urban complexes; a better description and understanding of the circulation and dominant mixing processes occurring in this particular region is clearly needed for a more effective management of the regional estuarine and shelf resources in the face of man's many conflicting uses of the water bodies. Fostered in part by increased environmental concerns, more adequate research funding, and the availability of new instrumentation and observational techniques, many new circulation and related physical studies have been undertaken in the last two decades, and a synthesis of both old and new material into a review of the regional estuarine and shelf circulation seems particularly appropriate at this time. While a few scientists have made important contributions in both fields and have thus helped to carry new ideas and techniques from one field into the other, basic research on problems concerning estuarine and continental-shelf circulation have evolved more or less independently in time, so that we will present here separate reviews of the historical and the modern ideas about estuarine and shelf circulation in the Middle Atlantic Bight. One important research objective in the 1980s will be to develop a better kinematic and dynamic description of the physical coupling between estuarine and shelf waters. Our present meager knowledge about the different physical processes that connect the shelf and estuary together prevent a more unified discussion.

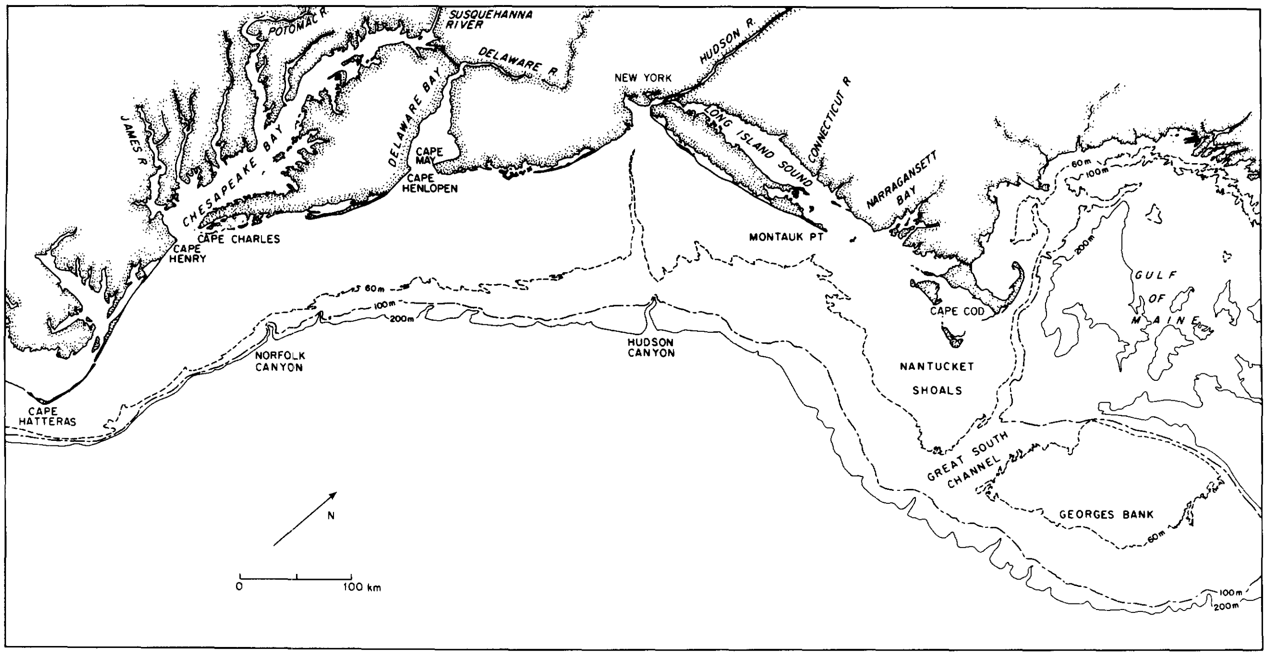

Figure 7.1 A topographic map for the Middle Atlantic Bight and a western section of the Gulf of Maine. The 60-, 100-, and 200-m isobaths are shown.

7.2 Estuarine Circulation in the Middle Atlantic Bight

We shall trace here the evolution of estuarine-circulation ideas. We will use the Middle Atlantic Bight estuaries as a focus because physical studies of these water bodies have played a major role not only in the development of the early concepts of the physics of estuaries but also in the recent refinement and reformulation of these ideas. Narragansett Bay, the Hudson River, Delaware Bay, and especially the Chesapeake Bay system—all have provided case studies from which significant advances have been made in our understanding of "coastal bodies of water having free connection with the open sea and within which sea water is measurably diluted by fresh water drainage" [Pritchard (1967a)].

7.2.1 The Development of Estuarine-Circulation Concepts

The idea that the introduction of fresh water into the sea can produce an oppositely directed, two-layer circulation has existed for a long time. Some early oceanographers understood that the lighter, fresher water spreads away from the source along the surface and the heavier, more saline water moves toward the source underneath. The kinematic details and the dynamics of this process have, however, remained elusive until recent years. On a scale as large as the Mediterranean Sea, early oceanographers correctly held that the circulations were driven by density differences. Two-layer flows observed at the Kattegat in the Baltic (Ekman, 1876), the Strait of Gibraltar in the Mediterranean Sea (Douglas, 1930), and the Strait of Bab el-Mandeb at the mouth of the Red Sea (Buchan, 1897) all reflect the balance between evaporation, runoff, and precipitation in the enclosed seas. On a smaller estuarine scale, the formulation of clear circulation ideas has been hindered by the difficulty of unraveling the different circulation components in a variety of estuarine geometries. In addition to oscillatory tidal currents and currents driven by density differences, there has been the possibility of "reaction currents" as suggested by F. L. Ekman (1876). These were countercurrents and undercurrents associated with the entrainment of ambient water by the discharge of a river into the sea. Although the idea of reaction currents seems to have been supported in Helland-Hansen and Nansen's (1909) discussion of the rivers entering the Norwegian Sea and in Buchanan's (1913) description of the flow off the mouth of the Congo River, F. L. Ekman's son, V. W. Ekman, showed that significant reaction currents were unlikely at the mouths of rivers (Ekman, 1899). The terms "reaction current," "induction current" (Cornish, 1898), "undercurrent" (Dawson, 1897), and "compensatory bottom current" (Johnstone, 1923) were often employed without detailed discussion of the physics. In some cases, these terms were used simply to refer to low-salinity water flowing seaward over more saline water flowing landward. In other cases, however, the usage harked back to F. L. Ekman's sense of a countercurrent associated with the river outflow jet.

Reaction currents were invoked to explain the two-layer circulation phenomena found in the early studies of the Middle Atlantic Bight estuaries. Although Harris (1907) correctly described the distribution of pressure surfaces in an estuarine situation, he interpreted Mitchell's (1889) observations on the Hudson River as illustrative of F. L. Ekman's countercurrent. Mitchell, in contrast, had interpreted the observed "underrun" of salt water below the outflowing fresher water as a response to a decrease in fresh-water discharge in the Hudson River. He calculated that "the surface of the river water would have to stand 2-1/2 feet above the ocean to prevent the salt water from running in along the bottom; and the sea-water would creep into the basin as soon as the head fell below this." If there were not mixing between the salt water and the fresh water, Mitchell's analysis would explain the movement of the density interface between the upper and lower layer and the position of the "neutral plane" where "inflow and outflow balance."

Although we now know there is significant mixing between the upper and lower layers in the Hudson, Mitchell's interpretation of his remarkably good current and density observations provided a reasonable explanation for the seasonal variation in salt intrusion. R. P. Cowles (1930) in his monograph on the Bureau of Fisheries' studies of the Chesapeake Bay waters noted that such behavior was not adequately explained by reaction currents. He wondered "why the undercurrent (as deduced by salt intrusion) moving in an ingoing direction is so marked during the winter months, when the discharge from the rivers is not ordinarily at its height?" That the surface layers were moving seaward was evident from the U.S. Coast and Geodetic Survey data collected by Haight, Finnegan, and Anderson (1930). They report 320 days of current pole measurements at the Chesapeake Bay mouth (Tail of the Horseshoe Lightship), where the mean current flowed out of the bay at 13 $cm\,s^{-1}$. At Thomas Point in the upper Bay, a 27-day record showed an 8-$cm\,s^{-1}$ mean flow seaward. Cowles mentioned many possible mechanisms for moving water and salt in Chesapeake Bay, and remarked on the difficulty in analyzing the resultant complexity. He did succeed in documenting the distribution and seasonal progression of temperature and salinity. Of particular note was the lateral gradient in salinity whereby the eastern side of the Bay was markedly saltier than the western side. Cowles ascribed this feature to the "fact that the deep-water channel which contains the most saline bottom water lies on (the eastern) side throughout most of its extent and to the fact that a large volume of fresh water from the rivers of the western shore presses the more saline water toward the eastern shore." He did not mention the possibility that the rotation of the earth played a role. Wells, Bailey, and Henderson (1929), who titrated the salinity samples for Cowles's surveys, did suggest that although the lateral gradient in salinity "had been ascribed to the fact that the principal rivers enter the Bay on its west side, the rotation of the earth may also be a factor."

Marmer (1925) thought that the greater nontidal surface flow (and greater ebb duration) along the western shore of the Hudson River was due to "the effect of the deflecting force of the earth's rotation." He noted that, at mid-depth, the flood-current velocities were greater on the eastern side of the river, and that the ebb velocities were greater on the western side. Marmer in his Hudson River study, and Zeskind and LeLacheur (1926) in their study of Delaware Bay, pointed out the decrease with depth in ebb-current duration in the estuary, and ascribed this decrease to the river discharge. Although they noted that the duration of flood may be greater than ebb near the bottom of the estuary, neither Marmer nor Zeskind and LeLacheur conveyed the sense that there is an internal, nontidal circulation present and that there is net up-estuary motion in the lower layer.

Haight's (1938) review of current measurements in Narragansett Bay contains little discussion of the nontidal flows. Although the U.S. Coast and Geodetic Survey's interest was to define and predict the tidal currents, Haight may have disregarded reporting the mean flows because the collected measurements were taken by a variety of methods under a variety of conditions, and because Narragansett Bay displayed such "irregularity of currents." Hicks (1959) later noted that Pillsbury's 1889 measurements [summarized by Haight (1938)] did show nontidal flow into the estuary in the lower layer.

Early current-measurement techniques did not allow easy determination of flow direction at depth. Mitchell's vertical profiles and Pillsbury's 5.5-day time series appear as notable achievements. Mitchell (1859) developed an apparatus to measure the "countercurrent" at depth in the Hudson River by modifying a device used to measure subsurface currents in European canals. Two floats (copper globes) were connected by a wire, one weighted to sink to a depth and act as a drogue. In order to reduce the errors resulting from the drag of the surface float, Mitchell added a third float attached to the surface float and weighted to have the same cross-sectional area. The attachment line of this additional float was connected to a reel to allow the two surface floats to separate freely. Mitchell argued that the original pair of floats would move at the mean of the surface and subsurface velocities and that the free float would move with the surface velocity. The Price current meter used by the U.S. Coast and Geodetic Survey had no provision for measuring direction. The common practice was to assume the current direction at depth corresponded to that indicated by the drift pole at the surface, a practice that could create substantial errors in nontidal-flow determinations in the estuary. Marmer's (1925) observations were successful because he employed a "bifilar direction indicator" in conjunction with the Price meter. This direction sensor, developed by Otto Petterson (Witting, 1930), consisted of a set of three vanes that were positioned at various depths and that transmitted their alignment to the surface by wires (Zeskind, 1926). Petterson also developed an internally recording current meter that could record speed and direction at 30-minute intervals for 2 weeks. The Petterson meter became available to the U.S. Coast and Geodetic Survey in 1925. The majority of current measurements reported in the cited survey reports on the Middle Atlantic Bight estuaries were made by current pole (often from anchored light ships) and Price current meters. Although Haight, Finnegan, and Anderson (1930) reported measurements made in the early 1920s by the U.S. Fisheries Commission employing Ekman current meters, their use does not seem to have been widespread.

The Coast and Geodetic Survey's collected current measurements in Long Island Sound were reported by LeLacheur and Sammons (1932). Again, they were primarily interested in tidal currents and they did not report the mean currents obtained from the long time series at the lightship. Prytherch (1929) released 500 drift bottles with drogues in his study of oyster-larvae transport and setting. With the 300 returns and a few Ekman and Price current-meter measurements, he deduced a net outflow from the Long Island Sound on the surface.

Important estuarine research was also being conducted elsewhere during the early decades of the twentieth century. Europeans and Canadians were active in the coastal regions where river runoff affects the regional general circulation. Palmén (1930) employed Bjerknes' (1898) solenoid method in a study of the wind-driven circulation in the Gulf of Finland. He demonstrated an oppositely directed two-layer flow and determined a wind-stress coefficient.

Jacobsen (1930) provided further details of the two-layer flow near the Kattegat through an analysis of current time-series measurements made from two lightships. Jacobsen also examined current and density measurements made in Randersfjord on the east coast of Jutland. From these data, which showed clearly the estuarine outflow and inflow (of the order of 10 cm s^{-1}), he calculated the surface slope along the axis of the fjord and determined coefficients of viscosity and mixing. He also considered the problem wherein a concentration of plankton was placed at the level of no net motion and allowed to disperse, illustrating the interaction of vertical diffusion and horizontal advection in such a two-layer system.

Following a suggestion by A. G. Huntsman that Watson's (1936) observations from Passamaquoddy Bay in the Gulf of Maine could be interpreted as a three-layer flow driven by tidal mixing, Hachey (1934) conducted a series of tank experiments in which he produced both two-layer and three-layer flows (figure 7.2). He concluded that

> the mixing of stratified water sets up dynamic gradients causing the following differential movements:
>
> (a) where, through the addition of fresh water at the mixing point, the mixed water is of a density which is less than that of the waters otherwise available for mixing, the mixed water is carried away from the mixing area in the upper layers, while a compensating current carries water to the mixing area in the lower layers; and
>
> (b) where the mixed water is of a density which is intermediate between the densities of the surface and bottom waters available for mixing, the mixed water is carried away from the mixing area at some intermediate level, and surface and bottom waters are carried to the mixing area to compensate for the waters entering into the mixing.
>
> A steady wind blowing towards the area of mixing is responsible for considerable modification of the above systems of currents. Such a wind seems to offer some resistance to the system outlined in (a), but considerably enhances a system of currents outlined in (b).

While the current measurements (in Digdeguash Harbor off Passamaquoddy Bay) offered by Hachey as an example of the three-layer flow may not be convincing because of their short duration and the uncertainties in density structure and in the strength of the wind-driven component, his conclusions from the tank ex-

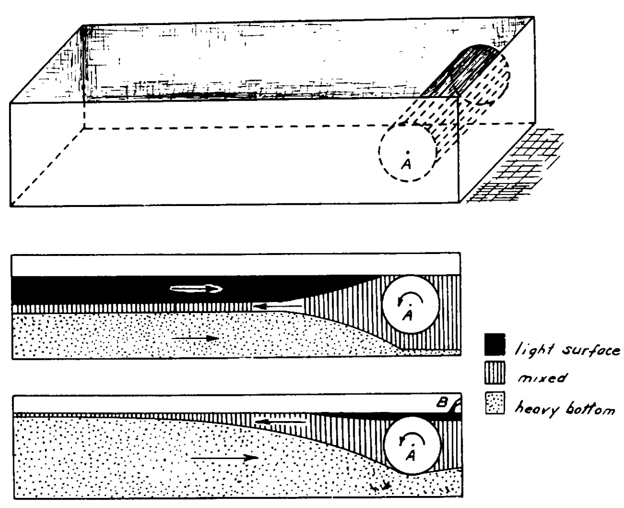

Figure 7.2 Diagram of Hachey's (1934) experimental approach (A), and resultant three-layer (B) and two-layer (C) circulation patterns. Mixing was provided by rotor *A* and fresh water was introduced by pipe *B*.

periments were correct and later substantiated by observations made in Baltimore Harbor. Hachey's work is especially significant because it was one of the first explanations of a density-driven circulation in stratified water generated by wind and tidal mixing.

In the years immediately following World War II, there was a marked increase in interest in the circulation of estuarine waters, caused in part by military needs and a heightened sense that the resources of the estuary and coastal waters were threatened by the nearby activities of man and should be protected.[1] Many of the papers from this period address the *flushing* characteristics of the estuary rather than the circulation per se. In addition to the Office of Naval Research's interest in basic research in the oceans, the Navy recognized a need for shallow-water studies to aid in mine warfare, amphibious warfare, and submarine-detection problems (Solberg, 1950). The Navy was also concerned with the possible environmental threat from nuclear submarine activity in bays, harbors, and estuaries. An indication of the scientific interest and talent dedicated to estuarine-circulation studies in the late 1940s is given in the proceedings (Stommel, 1950b) of the Colloquium on the Flushing of Estuaries held at the Massachusetts Institute of Technology in September 1950 and sponsored by the Office of Naval Research. The papers and discussion show not only that oceanographers were beginning to model the mixing processes in the estuary, but also that they were beginning to understand the possibility of an internal estuarine circulation.

The task at hand during the late 1940s was to determine the flushing mechanisms for estuaries. After Tully's (1949) extensive work on Alberni Inlet in British Columbia, much of the observational study was carried out on Middle Atlantic Bight estuaries. Ketchum (1950, 1951) sought to improve the tidal prism model whereby the sea water brought into the estuary on flood tide is assumed to mix completely with the water in the estuary. In addition, the water flushed out of the estuary on the following ebb is assumed lost to the system and does not reenter on the subsequent flood. Estuaries, however, do not mix completely on each tide. Ketchum therefore proposed to divide the estuary into successive volume segments the lengths of which were determined by tidal excursions. Within each segment complete mixing is assumed at high tide. Ketchum applied this concept to Tully's observations on Alberni Inlet and to his own study of Raritan Bay, New Jersey, and of Great Pond in Falmouth, Massachusetts, and achieved good agreement with the observed salinity distribution. Ketchum's success prompted Arons and Stommel (1951) to translate his segmented model into a continuous-mixing-length model. They produced a family of curves that were also successful in describing the salinity distribution in Alberni Inlet and Raritan Bay. The constant of proportionality relating eddy diffusivity to the tidal excursion and the tidal current amplitude differed, however, by an order of magnitude between the two estuaries. Pritchard (1965b) pointed out that these two treatments were applicable only to vertically homogeneous estuaries in which tidal mixing was sufficiently intense to eliminate vertical stratification. Stommel (1953b) later applied both Ketchum's model and that of Arons and Stommel to the Severn estuary, which has small vertical stratification. He showed that neither hypothesis worked for the Severn and mentioned that "it does not appear likely that any good purpose can be served at present by making *a priori* suppositions about the turbulent mixing process."

The obvious differences in the topography and salinity distributions in various estuaries led Stommel (1950b) to call for an estuarine-classification system employing differences in morphology and mixing processes as criteria. Stommel (1951) began the process with a classification scheme based primarily upon the "predominant physical causes of movement and mixing of water in the estuary," identifying river flow and tidal and wind mixing as the important processes. Pritchard (1952a, 1955, 1967b) and Cameron and Pritchard (1963) developed and refined Stommel's initial scheme. Hansen and Rattray (1966) later advanced the classification scheme by suggesting a two-parameter system that includes the stratification and the ratio of the net nontidal velocity at the surface to the river flow divided by the cross-sectional area of the estuary. An attempt at further refinement of these classification schemes has not been fruitful because of the difficulty in quantifying the parameter-selection process for a particular estuary. Many estuaries exhibit a variety of estuarine types.

In his discussion of estuarine classification, Pritchard (1952a, 1967a, 1967b) proposed the definition of an estuary quoted earlier as "a semi-enclosed coastal body of water which has a free connection with the open sea and within which sea water is measurably diluted with fresh water derived from land drainage." While this definition excludes inverse estuaries such as Laguna Madre, Texas, and San Diego Harbor, which are driven by evaporation, it is the most useful yet proffered because it sets the scale and the important elements controlling the characteristic estuarine circulation—lateral boundaries, the transmission of tidal energy and salt between the open sea and the estuary, and the introduction of sufficient fresh water to provide density gradients driving the currents. The Baltic Sea, for instance, would not be considered an estuary under this definition because its large scale renders the lateral boundaries less important to the kinematics and dynamics of water movement than they are in a true estuary.

One of the notable aspects of the development of estuarine-circulation concepts in the active decade following World War II was the extensive (and successful) use of laboratory models in deciphering mixing and transport processes. The first problems addressed with these models involved the simplest of estuarine types—the highly stratified or salt-wedge estuary. As physical oceanographers began to exchange ideas in meetings such as the 1950 colloquium at MIT, they became aware that the U.S. Army Corps of Engineers had been working with flumes and physical models for over 10 years at the U.S. Waterways Experiment Station in Vicksburg, Mississippi (Simmons, 1950). Among the earliest salt-intrusion studies were the Army Corps of Engineers' investigations of water-supply problems in the lower Mississippi River. The Army Corps of Engineers recognized a need for analytic help and in 1945 requested the aid of hydrodynamicists at the National Bureau of Standards "to investigate and establish the basic laws of similitude for models involving a study of density currents and the mixing of salt water and fresh water."[2] Keulegan provided this help and addressed many problems concerning the laboratory modeling of salt-wedge circulation. Keulegan (1949) produced salt wedges in flumes in which there was almost no mixing between the upper and lower layer. When he increased the flow of the upper layer, however, breaking internal waves formed on the fluid interface. Keulegan noted that, in this entrainment process, the waves only broke upward, carrying fluid from the lower layer to the upper layer.

Stommel and Farmer (1952) also examined the salt-wedge estuary with the aid of a laboratory flume. They showed that an abrupt widening in a channel can produce a stationary internal wave on the interface if the internal Froude number equals a critical value. This internal wave acts as a control on the outflow of the upper layer by restricting the thickness of the upper layer. Stommel and Farmer (1953) later noticed that if they added mixing to their flume, there was a point beyond which increased mixing has no effect on the outflow of the upper layer. Dyer (1973) explains that this "overmixing" mechanism is a result of the downward erosion of the density interface reaching the level where it restricts the compensating inflow in the lower layer. Model analyses were also conducted for wider, well-mixed estuaries such as Delaware Bay. Pritchard (1954a) studied flushing in the Army Corps of Engineers' Delaware model at Vicksburg, Mississippi, by employing dye as a tracer. Pritchard found that the eddy diffusivity was spatially scale dependent approximately in the proportion suggested by Stommel (1949).

The Chesapeake Bay Institute began a study of the moderately stratified James River in the summer of 1950 to examine the influence of the salinity and currents on the oyster seed-bed region in the middle reaches of the James River. The recent development of techniques for rapid sampling of currents and salinity from an anchored vessel allowed for the first time the collection of continuous detailed measurements for periods of 3 days or more. Current velocity was measured with a biplane drag (Pritchard and Burt, 1951), a modification of a method used by Jacobsen (1909) and apparently by Nansen (Witting, 1930), and salinity profiles were obtained with *in situ* conductivity and temperature sensors (Schiemer and Pritchard, 1957). These new observational tools were used to collect a data set sufficiently extensive in both time and space that meaningful temporal and spatial averages could be computed. This averaging procedure further minimized the (apparently low) variability due to local wind-driven currents and variations in the river flow that occurred during the sampling periods. Pritchard (1952b, 1954b) used this data set to evaluate the terms in the averaged salt-balance equation and to conclude that the horizontal advective flux and the vertical diffusive flux of salt were the most important in maintaining the balance. Pritchard (1956) also examined the momentum balance in the James River, determining the unknown terms in the equation of motion from the observations of the mean distribution of temperature, salinity, and current velocity. He evaluated the important Reynolds-stress terms and described the topography of the pressure surfaces, which sloped down toward the sea in the upper layer and down toward the head of the estuary in the lower layer. He found, as did Cameron (1951), that the cross-estuary pressure gradient and the Coriolis force were in approximate balance.

Rattray and Hansen (1962) used the James River observations and Pritchard's analysis to develop a theoretical steady-state circulation model for a moderately stratified estuary. Employing similarity transformations, whereby functional forms were assumed for the dependence of the stream function and salinity defect on the longitudinal position in the estuary, Rattray and Hansen reduced the two-dimensional partial differential equations governing the stream function and salinity defect to a pair of simultaneous ordinary differential equations. Under the conditions specified by Pritchard for the James River (in which the field accelerations and the vertical advective and horizontal diffusive fluxes were unimportant), and given the surface salinity distribution, Rattray and Hansen produced vertical profiles of salinity and velocity that matched those observed in the James River. Hansen and Rattray (1965) later relaxed some of the restrictive assumptions to retain the river-forced component of circulation.

Their work also found good agreement with the James River observations.

While this matching between these two theoretical treatments and Pritchard's analysis of the James River data has enhanced the attention paid to these studies, the value of these analytic models lies less in the agreement per se with observations than in the insight they provide into fundamental estuarine processes. These solutions to Pritchard's (1956) dynamic equations were the first to show clearly the interdependence of salinity and velocity fields in the estuary. While Agnew (1961) had considered two separate aspects of this interdependence in the free-convection part of the problem, Hansen and Rattray (1965) solved the coupled equations, including both the free-convection and forced-convection modes. Hansen and Rattray (1965) not only delineated the effects of fresh-water discharge and wind stress on the gravitational circulation in the James River, but also considered the interrelationships in an estuary such as the Mersey, which has a well-developed gravitational circulation despite the fact that tidal mixing nearly eliminates the vertical salinity gradient.

The concept and description of the internal circulation in a moderately stratified estuary evolved primarily from these studies of the James River data set. This circulation differed from the salt-wedge circulation, not only because the lower layer moved strongly toward the head of a moderately stratified estuary, but also because the transport in the individual layers was much greater than in the salt wedge. Near the mouth of a moderately stratified estuary, the upper-layer net (nontidal) transport can be an order of magnitude greater than the river flow entering the estuary. In describing the driving mechanism for this internal circulation, Pritchard (1967b) stated:

> It has been attributed to the increased potential energy of the system which follows from increased exchange between the fresh-water and saltwater layers. More accurately, tidal mixing produces horizontal density gradients of increased strength, which in turn produce horizontal pressure gradients of sufficient magnitude and extent to maintain the relatively higher velocities even in the face of increased eddy friction. Tidal mixing is responsible for both the increase in potential energy and the distribution of potential energy within the estuary.

7.2.2 Recent Developments in the Study of Estuarine-Circulation Processes

While the contributions of Pritchard, Rattray, and Hansen represent significant advances in our understanding of estuarine-circulation processes, fundamental questions remain as to the nature of the transport of salt and momentum, the role of the wind in the transport processes, and the effects of topography in producing both order and disorder. In spite of an improved ability to attain spatial coverage and resolution with modern instrumentation, our ability to describe and model estuarine physics is still limited by inadequate parameterizations of friction and turbulent mixing. Longer current-meter records are showing that the current variability due to wind forcing is more complex than previously thought. As more detailed information on the circulation becomes available, there is a growing conviction on the part of estuarine investigators that the variations in the lateral direction are significant in the dynamics, and that bottom topography can generate both secondary flows and residual circulations.

Three observational methods have been employed to separate and examine mixing processes in an estuary: (a) the evaluation of terms in the temporally and spatially averaged salt-balance equation; (b) the direct measurement of turbulent fluctuations in velocity and salinity; and (c) the observation of dispersion by an introduced tracer. The first method is the analytic technique employed by Pritchard (1952b, 1954b) on the James River data set. His conclusion that the horizontal advective flux of salt and the vertical diffusive flux are the dominant terms is based on an analysis that assumes lateral homogeneity. For estuaries such as Delaware Bay, which can exhibit vertical homogeneity but have lateral gradients in salinity and velocity, Pritchard (1955) suggests that, by analogy with the James River, the dominant salt-flux terms are probably the lateral-diffusive and the longitudinal-advective terms. These analyses involved tidally and spatially averaged values of salinity and velocity, but the averaging process is not explicitly developed in the salt-balance equation. Pritchard (1958) begins the rigorous averaging of the three-dimensional salt-balance equation, expressing salt and velocity variables as sums of time-mean values and deviation terms. Bowden (1963) and Cameron and Pritchard (1963) further decompose the variables into a time mean, a turbulent fluctuation, and a single oscillatory term varying sinusoidally over the tidal cycle. Bowden employs this decomposition to examine the effect of vertical shear on the longitudinal transport of salt in a laterally homogeneous estuary. He finds that, for the Mersey River, there are occasions when the advective flux of salt out of the estuary (driven by the river discharge) is approximately balanced by the transport associated with the vertical shear in velocity and vertical variations in salinity. On other occasions, Bowden finds that the up-estuary transport is shared between the "shear effect" and the transport arising from the correlation between the harmonically varying terms of the depth-mean velocity and salinity. There are also times when this tidal-correlation term dominates and times when the upstream and downstream salt transports do not balance.

Okubo (1964) has carefully examined the averaging process for an estuary with lateral as well as vertical variations, specifying the assumptions under which his salt-balance equation is appropriate. He uses the salt equation averaged over the cross section in a successful analysis of measurements made at the Delaware estuary model at the U.S. Waterways Experiment Station. Hansen (1965) also considers variations over the cross section of the estuary. He decomposes the cross-sectional mean variables into a mean, a harmonic tidal variation, and a turbulent fluctuation. For the Columbia estuary, which has a large river flow, a large tidal range, and a weak gravitational circulation, Hansen finds that the advection of salt driven by the river discharge is balanced primarily by fluxes associated with the correlation of velocity and salinity fluctuations of the tidal period and with the shear effect. Fischer (1972) argues that for the Mersey, the salt flux associated with the lateral shear is not only larger than that associated with the vertical shear, but that it is dominant. He proposes decomposing the deviations from the cross-sectional mean into variations in the vertical and lateral directions. Fischer's conclusions for the Mersey stand in contrast to the analysis by Bowden and Gilligan (1971) of Mersey observations made in the reach where density currents are significant. There may be agreement for the reaches seaward of this region. While Fischer's point that the lateral shear can make a significant contribution to the longitudinal transport of salt is well taken, his estimates of the terms in the salt-flux equation are based primarily on parameterizations of the dispersion coefficients and not on direct computations using the salinity and velocity observations in the manner of Pritchard, Bowden, Okubo, or Hansen. Dyer (1973) states:

> So far it is not possible to define precisely which are the dominant factors since different investigators have used slightly different methods of analysis; they split up their components in a variety of ways with certain implicit assumptions. Consequently, the results of differences in tidal response and topography between estuaries are not clear.

The approach of averaging the salt-balance equation does seem to offer a promising means of attaining an explicit separation of the flux components. Dyer (1973) suggests combining Hansen's (1965) and Fischer's (1972) schemes, and applies (Dyer, 1977) the full set of terms to a salt-wedge, a partially mixed, and a well-mixed estuary. He finds that for Southampton Water, a partially mixed estuary, the salt flux associated with the vertical shear and the lateral shear are of the same order. It is clear that great care is required to avoid dependence on the observational scheme and method of data handling. In light of the recently observed wind-driven variability in estuaries and the importance of topographic effects, proper evaluation of this approach will require long record lengths, good spatial coverage and resolution, and shrewdness in averaging procedures. Rattray (1977) calls for both better methods of integrating the governing equations in conjunction with field programs and more extensive and elaborate field observations.

While averaging and evaluating the various terms in the salt-balance equation provides insight into the spatially integrated mixing processes in the estuary, the direct measurement of turbulent fluctuations provides a unique look at mixing on a small scale. This complementary method is particularly suited for the determination of the source(s) of mixing, about which little is now known. The various roles of wind stress, shear at the pycnocline, bottom stress, and surface and internal waves in providing the turbulent mixing of salt have not yet been evaluated. Bowden (1977) reviews turbulence measurements made in estuaries [including a noteworthy early attempt by Francis, Stommel, Farmer, and Parson (1953)] and the subsequent attempts to parameterize the observed fluctuations for the construction of models. The measurement of turbulent velocity fluctuations has required many innovative techniques. Bowden and Fairbairn (1952) mounted two Dodson-propeller current meters on a rigid stand on the bottom of the Mersey estuary. This device employed a spring-loaded propeller, which enabled a rapid response to the turbulence. Bowden and Howe (1963) were the first to employ an electromagnetic flow sensor [developed earlier by Bowden and Fairbairn (1956)] to measure turbulent fluctuations in an estuary. Many of the subsequent turbulent-velocity measurements were made in estuaries tributary to the Chesapeake Bay. Cannon (1971) used a biaxial current meter (Cannon and Pritchard, 1971) to measure intermediate-scale turbulence from a tower erected in the Patuxent River. Seitz (1973) also measured turbulence in the Patuxent River, using an acoustic Doppler-shift current meter (Wiseman, Crosby, and Pritchard, 1972). He showed the approach to isotropy of the three Cartesian velocity components at high wavenumber, and also showed the spectral distribution of horizontal shearing stress, which reaches a peak at intermediate wavenumbers. Pronounced intermittency in the Reynolds stresses in the Choptank River was reported by C. M. Gordon (1974). J. D. Smith (1978) has developed a profiling system which is particularly suited for measuring turbulent fluctuations of temperature, conductivity, and velocity in an estuary. With this instrumentation, Gardner and Smith (1978) investigated mixing events in the Duwamish salt-wedge estuary in Washington that are apparently triggered by a hydraulic jump that occurs at a sharp change in river depth.

Tracking the dispersion of an introduced dye tracer provides a third method for examining mixing processes in an estuary. Pritchard and Carpenter (1960) developed a technique for detecting a concentration of 0.04 parts per billion of Rhodamine dye. The three-layer circulation of Baltimore Harbor was discovered through the use of this dye-tracer technique. An application of particular interest is the examination of the shear effect by Wilson and Okubo (1978) in the York River, off Chesapeake Bay. They employed Okubo's (1967, 1969) theoretical methods to analyze the dispersion of a dye release in the lower layer of the York estuary. They provide a model for separating the longitudinal dispersion in a stratified estuary due to horizontal turbulence and to the interaction of vertical shear with vertical mixing. They also include the modifications to the shear effect caused by the nontidal upward advection.

Although early investigators showed a keen awareness of the effects of strong wind forcing on estuarine circulation, and they often invoked wind effects to explain discrepancies that arise in interpretations that ignore wind driving, the significance of wind-driven circulations in estuaries has only been recently discovered through the analysis of long current observations. Pickard and Rodgers (1959) show a wind-induced shift in the mean velocity profile in Knight Inlet, British Columbia. Hansen and Rattray (1965) suggest that even a small wind stress could have a marked influence on the gravitational circulation. Weisburg and Sturges (1976) were among the first, however, to examine the wind transport in a partially mixed estuary using long-term current-meter data. They show, using month-long current measurements from Narragansett Bay, that wind transients can easily dominate the longitudinal flux of water in an estuary and that a proper separation of the gravitational circulation is difficult with short records (Weisberg, 1976a). Weisberg (1976b) has developed a stochastic model for the wind-driven longitudinal flow at one position in the Providence River in Narragansett Bay. Farmer and Osborn (1976) describe the wind circulation in Alberni Inlet. Up-estuary winds can reverse the current in the upper low-salinity layer and cause a deepening of this layer near the head. Farmer (1976) presents a simple model for the freshwater thickness in the upper reaches of the inlet, and produces an accurate simulation of the wind-driven behavior.

A year-long series of current measurements made in the Potomac River estuary led Elliott (1978) to the discovery that the estuarine circulation was not only affected by local wind forcing, but also by sea level in the Chesapeake Bay proper. This nonlocal forcing is examined by Wang and Elliott (1978), who find that the nonlocal forcing extends to the continental shelf. The dominant sea-level fluctuations in the Chesapeake Bay have a period of 20 days and are the result of up-estuary propagation of coastal sea-level fluctuations. Local winds operate on a shorter time scale, driving seiche oscillations in the bay at a period of 2.5 days. Wang (1979a,b) has examined this wind driving further, and finds that the predominant current fluctuations in the lower Chesapeake Bay are barotropic.

The increasing evidence for wind control on time scales of 10 days or less leads to speculation on the role of wind mixing versus tidal mixing in providing the energy source for the gravitational circulation. While we do not have sufficient evidence at present to decide this question, investigators are beginning to reveal both the mode and details of the topographic effects on the tidal mixing process. The oscillatory movement of the tides acting on shoreline irregularities and complex bottom topography is known to increase the longitudinal dispersion in estuaries (Pritchard, 1953; Holley, Harleman, and Fischer, 1970; Okubo, 1973). Sugimoto (1975) and Zimmerman (1978) describe the production of residual vortices by propagation of the tidal wave over a complicated topography. Ianello (1977) and Zimmerman (1979) remind us that, for transport processes, careful consideration of the Stokes drift must be given. The inherent errors and logistical difficulty of Lagrangian current measurements as yet leave us with Eulerian measurements as the only means of spatial and temporal coverage. Rattray's (1977) call for elaborate and extensive measurement programs should be repeated if we are to consider the measurement of the Stokes drift by Eulerian means.

The generation of secondary flows by bends in rivers is well known to fluid dynamicists and geologists. Secondary flows in estuaries that have stratification and tidal oscillation, however, are less well understood, partly because observational evidence is scanty. Dyer (1977) outlines the expected cross-estuary flow pattern for various degrees of stratification and shows, as does Stewart (1957), that the field-acceleration terms cannot be neglected in the lateral dynamic equation when there is curvature in the estuary.

Episodic tidal-mixing events may occur not only on a time scale of the semidiurnal tide, but also on a scale of the fortnightly variation in tidal range. Haas (1977) reports observations from the lower York River and Rappahannock River on the Chesapeake Bay and suggests that, in these rivers, the increase in tidal-mixing energy from neaps to springs provides sufficient increase in tidal mixing to eliminate the vertical stratification. Cannon and Ebbesmeyer (1978) and Cannon and Laird (1978) describe fortnightly salinity intrusions in the fjordlike Puget Sound estuary in Washington. These events are associated with the large spring tides over the entrance sill.

Garvine (1977) shows that lateral fronts in estuaries may be important to both vertical and lateral mixing.

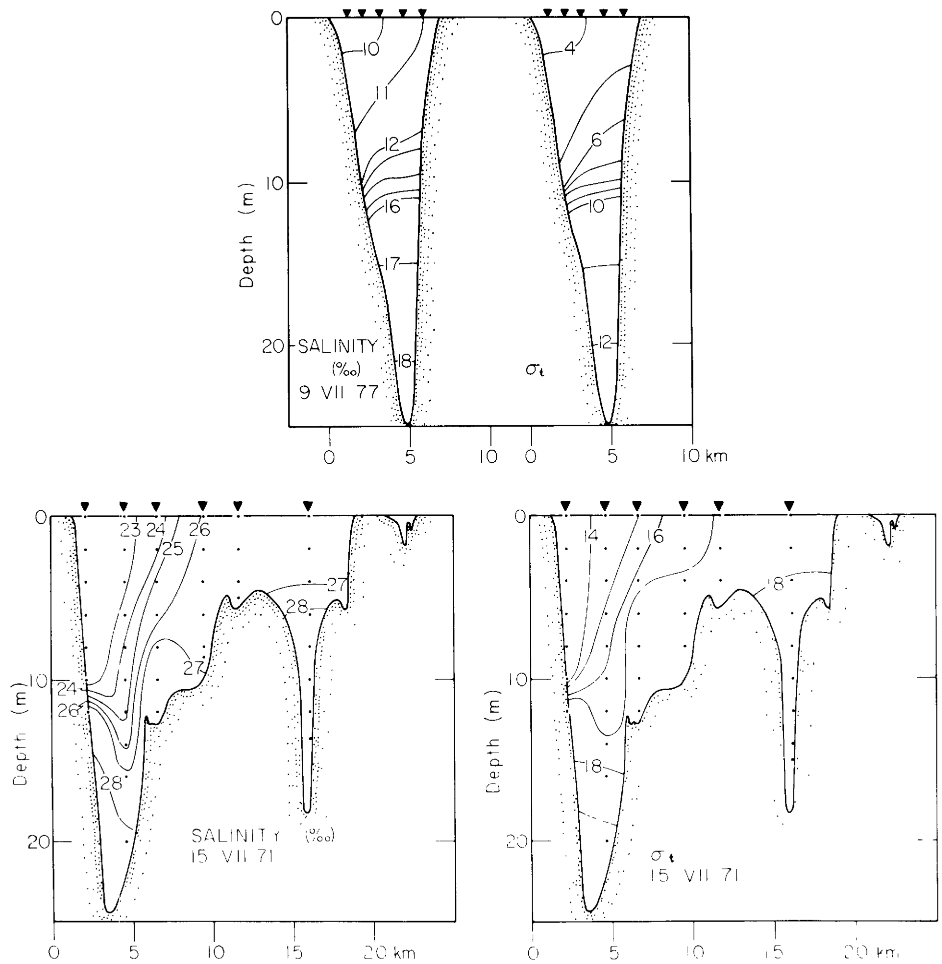

Figure 7.3 Salinity and density (σ_t) distributions at two cross sections of the Chesapeake Bay. Top section is located near Annapolis, Maryland, approximately 220 km up the estuary from the mouth, and bottom section is located at the mouth of the Bay between the Virginia Capes.

These fronts may be tidally time dependent or occur when the pycnocline breaks the surface, as in lower Chesapeake Bay. Figure 7.3 illustrates salinity and density distributions at two positions in Chesapeake Bay; one section (top) is near Annapolis, Maryland, approximately 220 km up the estuary from the mouth, and the other section (bottom) is between the Virginia capes at the mouth of the Bay. The cross-estuary tilt of the pycnocline is evident in the Annapolis section. If this cross-estuary tilt is approximately in geostrophic balance, the increase in transport in the gravitational circulation toward the mouth of the bay requires a corresponding increase in tilt. In the lower Chesapeake Bay, the tilt increases to the point where the pycnocline breaks the surface, often appearing as a series of strong lateral fronts. This observed increase in tilt is probably the combined result of the increase in the geostrophically balanced gravitational flow, the addition of fresh water by rivers on the western side of the bay, and by the widening of the bay in the lower reaches.

The salinity and density sections shown in figure 7.3 serve to illustrate the inherent three-dimensionality of the flow near the mouths of estuaries. This three-dimensionality and complexity near the mouth often makes it difficult to formulate realistic boundary conditions for numerical circulation models of the estuary. While the estuary does not often strongly affect the circulation on the adjacent continental shelf, the estuary often dominates the flow in the mouth and in the nearshore regions.

7.3 Continental-Shelf Circulation

We shall discuss in this section some ideas and observations about the general circulation over the continental shelf in the Middle Atlantic Bight. Bumpus (1973) and Beardsley, Boicourt, and Hansen (1976) have presented recent reviews on the circulation within the Middle Atlantic Bight. Bumpus (1973) describes some of the historical ideas about the Middle Atlantic Bight circulation and gives a summary interpretation of the large amount of surface-drift bottle and sea-bed-drifter data acquired during the 1960s over the eastern United States continental shelf. In the 1970s, moored arrays of self-contained current meters and other *in situ* instrumentation have been deployed in the Middle Atlantic Bight, and Beardsley, Boicourt, and Hansen (1976) present some of the preliminary results from these new field programs. In the 4 years since that review, longer current-meter records have been obtained and other descriptive and theoretical advances have occurred, making it seem both worthwhile and appropriate for us to attempt here to update the preliminary physical picture presented in 1976.

We shall begin with a brief physiographic description of the Middle Atlantic Bight and then present a review of the early observational work and ideas about water structure and the general circulation in the Middle Atlantic Bight. This review is presented both for completeness and to give the reader a sense of the origin and evolution of key ideas and observational methods used to study the shelf circulation. We shall next describe the nature and structure of atmospheric forcing over the Middle Atlantic Bight because the early moored-array work demonstrated that much of the subtidal current variability observed in the Middle Atlantic Bight is directly wind driven. We shall next describe what is known about the temporal and spatial structure of the wind-driven subtidal transient circulation on both the synoptic (2-to-10-day) time scale and the longer monthly time scale. The observed mean current field and ideas about how it is driven and maintained will be discussed at the end.

7.3.1 Physiographic Setting

Uchupi's (1965) bathymetric map shows that the shelf topography within the Middle Atlantic Bight is relatively simple and smooth in comparison to the more complex topography within the Gulf of Maine and Scotian Shelf region. The depth within the Middle Atlantic Bight generally increases in a monotonic fashion from shore out to the shelf break. The depth of the shelf break decreases from about 150 m south of Georges Bank to about 50 m off Cape Hatteras. The width of the shelf from shore to shelf break is generally about 100 km except near Cape Hatteras, where the shelf becomes very narrow (about 50 km), and near New York, where the New Jersey and Long Island coasts form a corner region making the shelf there about 150 km wide. Both the mean depth and cross-sectional area of the shelf decrease roughly by a factor of two from the New England shelf to off Cape Henry. The continental slope is indented by many submarine canyons, but only a few penetrate up onto the outer shelf. Several drowned river channels partially cross the shelf, the most notable being the Hudson River Channel off New York (see figure 7.1).

Milliman, Pilkey, and Ross (1972) have mapped the superficial sediments over the eastern United States continental margin and find the Middle Atlantic Bight to be covered mostly with medium-sized sand. Finer-grained sediments are found in a large region southwest of Nantucket, near the major estuaries, and generally seaward of the shelf break. A wide spectrum of small-scale morphological features exists over much of the shelf, ranging from wave-formed ripples 10 to 15 cm long and 1 to 10 cm high up to large-scale ridges 2 to 4 km long and up to 10 m high. Intermediate-scale features such as sand waves of varying size are frequently superimposed on the larger-scale features. The topography in the transition region between estuary and inner shelf is complex and most estuaries within the Middle Atlantic Bight have at least one relatively deep channel connecting the estuary and shelf. These larger-scale topographic features can influence currents through both topographic steering and generation of horizontal eddies, while the smaller-scale features can exert a significant form drag on the flow. More detailed descriptions of the Middle Atlantic Bight bottom topography, superficial-sediment distribution, and ideas about the formation of these features are given by Emery and Uchupi (1972), Swift, Duane, and McKinney (1973), Swift et al. (1976), Freeland, Swift, Stubblefield, and Cok (1976), and Freeland and Swift (1979).

7.3.2 Early Development of Ideas about the Shelf Circulation

It was considered accepted knowledge before 1915 that a rather sharp transition zone existed near the shelf break between the generally cooler and fresher "coastal" water found over the shelf in the Middle Atlantic Bight and the generally warmer and more saline "Gulf Stream" water found offshore.[3] The textbooks and ocean atlases of this early period [e.g., Findlay (1853), Maury (1855), and the current chart of the U.S. Navy published by Soley (1911)] showed the coastal water to be generally moving slowly toward the southwest along the shelf from Nova Scotia to Cape Hatteras. The low temperature and salinity of the coastal water suggested a northern origin, and Verrill (1873), among others, emphasized that the coastal currents supported a boreal littoral fauna rather than the warm-water fauna characteristic of the Gulf Stream. The cold coastal water had been mapped as far north as Newfoundland and most oceanographers like Libbey (1891, 1895) and Sumner, Osburn, and Cole (1913) believed that the Labrador Current flowed along the coast from the Grand Banks past Nova Scotia and the Gulf of Maine into the Middle Atlantic Bight and perhaps even as far south as Florida. This belief was modified when Schott (1897) and Dawson (1913) showed, using direct-current as well as temperature and salinity measurements, that the outflow of the Gulf of St. Lawrence via the Cabot Straits is the primary source of coastal water on the Scotian Shelf. The British Admiralty (1903) charts show this coastal water flowing toward the southwest into the Gulf of Maine at Cape Sable, where the current either turned northward toward the Bay of Fundy, or became too diffuse to determine from the mariner reports.

In 1912, H. B. Bigelow began a remarkable series of cruises that provided the first comprehensive description of the hydrography, circulation, and biology of the Middle Atlantic Bight and Gulf of Maine region. Bigelow had first gone to sea as a college undergraduate with Alexander Agassiz in 1902 (Schlee, 1973), and after finishing his doctorate at Harvard in 1906, he joined Agassiz as a research assistant at the Museum of Comparative Zoology, where he spent much of his time describing and classifying jellyfish collected on Agassiz's expeditions. In 1908, an ailing Agassiz directed Bigelow to conduct a short cruise of his own across the continental shelf to collect animals from the Gulf Stream, which Bigelow did aboard the Bureau of Fisheries' 90-foot schooner *Grampus.* Agassiz died in the summer of 1910, and Bigelow spent the next year working on jellyfish at the Museum and reading about the research being conducted in the eastern North Atlantic by Scandinavian scientists under the guidance of J. Hjort, the Director of the Norwegian Board of Sea Fisheries. Then Sir John Murray visited Harvard in 1911 and convinced Bigelow to leave the laboratory for a time and launch his own expedition, which he eagerly did the following summer aboard the *Grampus* (Schlee, 1973).

It seems clear that Bigelow, in developing his own field program, was strongly influenced by both Hjort's systematic approach to oceanographic research (see Schlee, 1973) and several key technological advances made by the Scandinavians in the period 1900–1910 (see chapter 14). Knudsen (1901) had prepared tables for conveniently calculating salinity and density at atmospheric pressure (σ_t) from values of temperature and chlorinity, Ekman had developed a mechanically recording propeller-type current meter [see von Arx (1962) for a description] that could be used from an anchored ship, and Nansen had perfected a practicable reversing water sampler with an attached thermometer. Equipped with these new tools, plus a variety of improved biological and geological sampling gear, and sponsored by the Bureau of Fisheries and the Museum of Comparative Zoology, Bigelow and his coworkers set sail on the *Grampus* in July 1912 to study the hydrography, currents, and biology of the Gulf of Maine. Bigelow conducted a similar research cruise in the next summer.

In 1915, Bigelow published his first tentative chart (shown here in figure 7.4) of the summer surface circulation for the Gulf of Maine and the Middle Atlantic Bight region, as inferred from the July 1913 cruise. Bigelow stated that "the combined evidence of the various records of ocean currents, our own included, points to the conclusion that the dominant drift over the continental shelf south of New York is to the southwest; and this is certainly the prevalent opinion of practical navigators and hydrographers" [p. 232]. His chart suggested the importance of runoff from the major estuaries within the Middle Atlantic Bight in both the surface salinity and current patterns. Bigelow also speculated that the cold bottom water found in the Middle Atlantic Bight was formed locally in the previous winter and was essentially static and not advected into the Middle Atlantic Bight from the east. He (1922) found further support for this idea in the August 1916 data. While the idea of a mean near-surface drift toward the southwest in the Middle Atlantic Bight has been confirmed by more modern measurements, the concept of the cold bottom water as static was clearly refuted when direct-current measurements began in the 1970s.

Bigelow was primarily interested in the Gulf of Maine during this period, however, and after a brief interruption due to World War I, he resumed his field work and began to focus more on the circulation there. He began to release surface drift bottles along strategic sections within the Gulf of Maine and also experimented with E. Smith with the Scandinavian method for geostrophic-current computation.[4] In 1927 Bigelow's monograph on the physical oceanography of the Gulf of Maine was published by the Bureau of Fisheries. Using hydrographic data and geostrophic computations as well as current information inferred from the movement of fish eggs and larvae and drift bottles, Bigelow developed a rather accurate conceptual model of the general circulation of the Gulf of Maine on a seasonal time scale, which has become the foundation for all subsequent work in this region. He described the springtime formation of a counterclockwise circulation around the basin (called the Gulf of Maine gyre) and a clockwise circulation around Georges Bank (the Georges Bank gyre). He recognized that slope water characterized by $S \geq 35‰$ penetrated through the Northeast Channel into the deeper basins of the Gulf of Maine and that this water mixed with very fresh shelf water from the Scotian Shelf and farther north to form the intermediate salinity water found in the Gulf. Bigelow's schematic near-surface circulation diagram (1927, p. 973) showed that at least during the summer (when drift-bottle returns were highest), some shelf water flowed westward past Nantucket Shoals into the Middle Atlantic Bight.

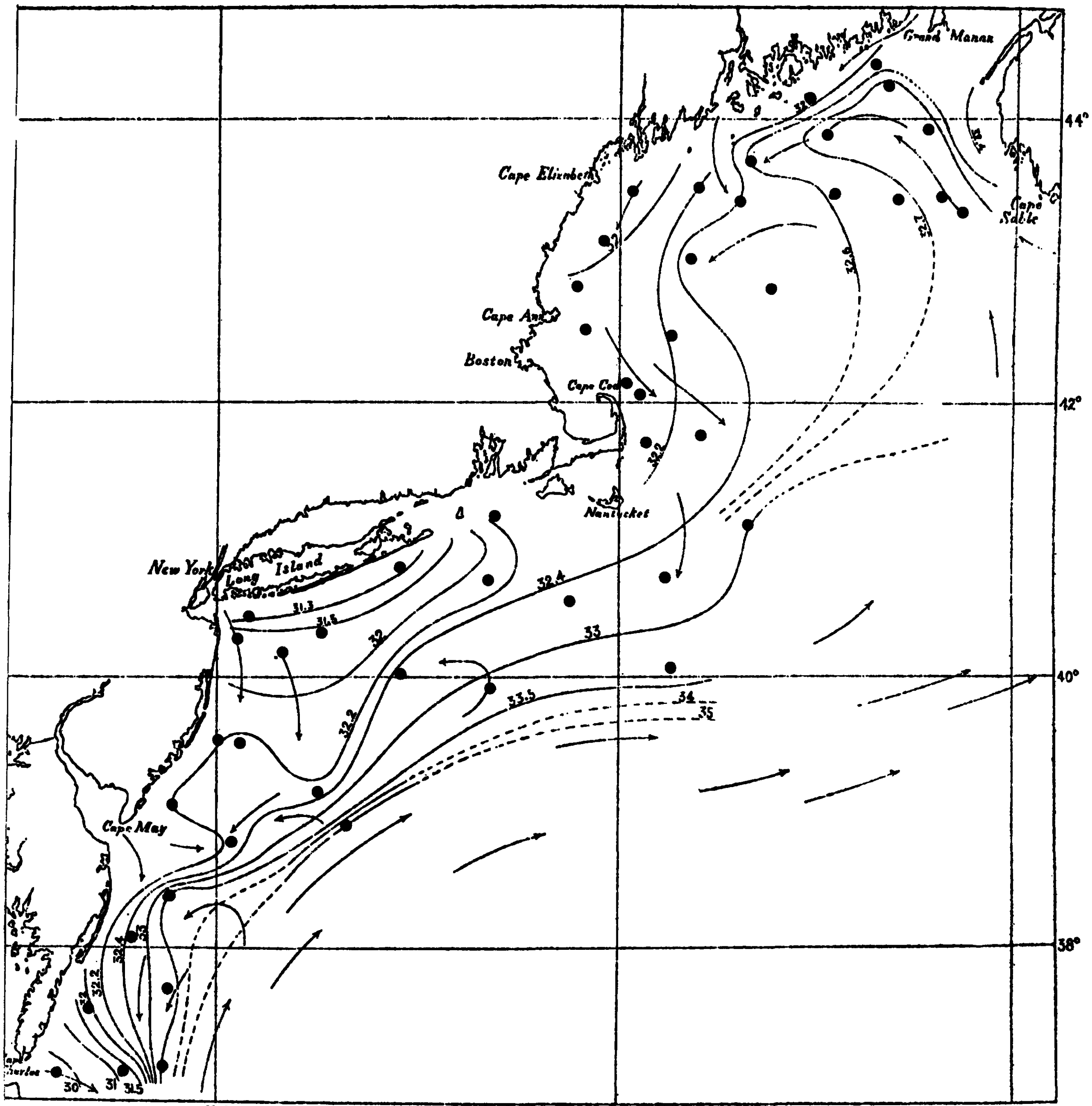

Figure 7.4 Surface circulation map for July 1913 published by Bigelow (1915). Surface salinities are shown and dots have been added to show hydrographic station locations.

Summarizing their past work and incorporating some new measurements made aboard the *Atlantis*,[5] Bigelow (1933) and Bigelow and Sears (1935) produced the first complete description of the seasonal temperature and salinity fields within the Middle Atlantic Bight. They found that vernal warming and fresh-water runoff built a strong stratification during the late spring and summer months, which was subsequently destroyed in the fall and early winter by surface cooling and winter storms. Bigelow and Sears recognized that shelf water represented a mixture of continental runoff and the more saline slope water and documented the basic structure of the transition zone between these two water masses. The transition from shelf to slope water often occurred as a sharp outward-sloping front located near the shelf break during winter, while the front was less distinct in summer because of the development of a seasonal thermocline in the adjacent slope water. Large temperature and salinity gradients still persisted in the offshore direction below the seasonal thermocline on account of a band of cold, low-salinity shelf water that was located near the bottom on the outer shelf and was described by Bigelow (1915, 1922, 1933) as a remnant from the previous winter cooling. Bigelow incorrectly visualized an essentially static pool of cold bottom water extending from south of Long Island to Cape Henry that was entirely surrounded by warmer water and persisted without replenishment through the summer.

In the late 1930s, Bigelow's personal research returned to fish and he did not write further about coastal circulation per se. Bigelow and C. Iselin did encourage a 3-year interdisciplinary field study of the Georges Bank region and its high biological productivity (Schlee, 1978), and, although it was stopped early in 1941 by World War II, this study did produce a number of biological and ecological papers, including one on ecosystem modeling (Riley, Stommel, and Bumpus, 1949). Based on his own work on slope water and the Bigelow-Sears picture of the Middle Atlantic Bight hydrography, Iselin (1939b) stated without discussion that "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." The idea that the geostrophic balance represented a driving mechanism was apparently a common misconception. Iselin clearly believed that the density distribution over the shelf and slope was the principal driving mechanism of the shelf circulation. The maximum horizontal density gradients occurred in the frontal zone near the shelf break, so with an assumed level-of-no-motion near the bottom, as suggested by Bigelow (1915, 1922, 1933), the surface geostrophic current would be a maximum, and directed toward the southwest along the shelf break. Iselin (1939b, 1940b) did correctly point out that the generally observed increase of salinity with depth over the shelf in the Middle Atlantic Bight implies an offshore motion near the surface and an onshore flow at depth.

The first dynamic model for the Middle Atlantic Bight circulation was published in *The Oceans* by Sverdrup, Johnson, and Fleming (1942). The circulation scheme shown in figure 7.5 is taken from *The Oceans* and indicates a drift of coastal (meaning shelf) and slope water along the continental margin towards the southwest in the correct sense. According to Sverdrup, Johnson, and Fleming (1942, pp. 677–680), precise geodetic leveling experiments conducted in the early 1930s indicated that mean coastal sea level rose between Cape Hatteras and Cape Cod by some 10 cm. The north-south gradient of mean atmospheric pressure was known to be small enough that oceanographers believed that these measurements indicated a *real* northward rise in the absolute sea-surface topography. Since the Gulf Stream presumably did not run uphill, and Dietrich (1937) had "showed" that the northward surface slope was *not* caused by a northward decrease in mean density along the slope, Sverdrup inferred that the sea surface had the profile labeled 2 in figure 7.5, which would imply a southwestward geostrophic current over the shelf with a maximum near the shelf break as argued by Iselin (1936, 1939b). Sverdrup stated that "a current to the south must also flow over the

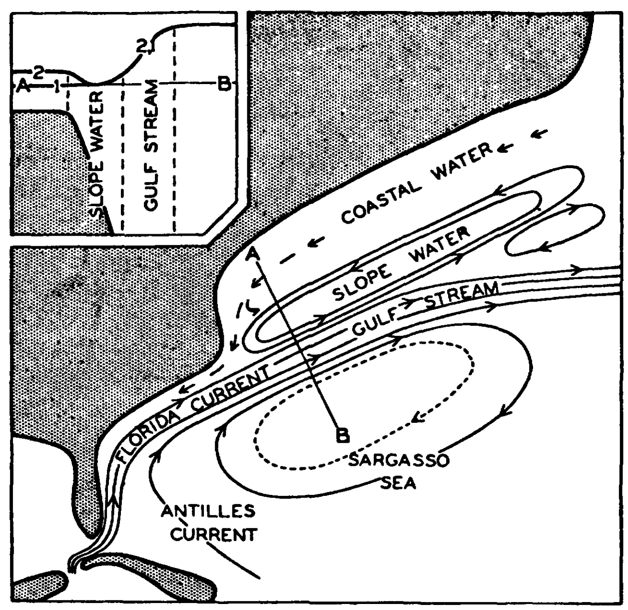

Figure 7.5 Schematic representation of the character of the Gulf Stream, taking results of precise leveling into account. Inset: Profiles of the sea surface along the line A–B. Profile 1 derived from oceanographic data only; Profile 2, from these data and the results of precise leveling. [Circulation scheme given by Sverdrup, Johnson, and Fleming (1942).]

shallow portion of the shelf where it flows downhill and where the balance of forces is maintained by the effect of friction" [Sverdrup, Johnson, and Fleming (1942, p. 678)]. Sverdrup speculated that this surface topography pattern was caused by large-scale wind forcing but his reasoning was vague. Even though the accuracy of the geodetic leveling has been disputed by Sturges (1968) and others, recent circulation models also invoke a mean alongshore pressure gradient. This point will be discussed again below (and see chapter 4).

Haight (1942) published the first long-term *surface*-current observations made in the Middle Atlantic Bight and Gulf of Maine region. The U.S. Coast and Geodetic Survey had a 30-year-long cooperative program with the Lighthouse Service and the Coast Guard to measure surface currents using the current drift-pole technique at lightships and other stations on the shelf. Haight presented quite accurate charts for the tidal currents and summary charts for the nontidal or mean and wind-driven currents. This tidal-current information and other direct measurements made in the estuaries and harbors form the basis for the current roses found on today's navigation charts.

World War II stopped active research on the Middle Atlantic Bight and Gulf of Maine, and although a number of useful instruments like the bathythermograph (BT) and Loran A were developed and perfected, and many BT profiles were taken over the shelf, most oceanographers were busy with defense-related research, and active work on the Middle Atlantic Bight did not resume until the late 1940s. Spilhaus and Miller (1948) modified the BT to obtain discrete water samples while ascending, and Spilhaus, Ehrlich, and Miller (1950) and Miller (1950) then used this new instrument to examine the shelf-slope water front south of New England. Miller (1950) found evidence for significant mixing across the deeper σ_t-surfaces near the shelf break, which he attributed to internal wave breaking in the frontal zone. Ford, Longard, and Banks (1952) found narrow filaments of relatively cold and fresh water along the shoreward edge of the Gulf Stream north of Cape Hatteras, and Ford and Miller (1952) correctly surmised that this water was, in fact, shelf water from the Middle Atlantic Bight entrained along the edge of the Gulf Stream near Cape Hatteras.

In 1950, the National Lead Company began to dump acid-iron waste from barges in the New York Bight and a number of oceanographers were asked by the National Research Council to examine the environmental effects and estimate the flushing time for the New York Bight. The results of this work were reported by Redfield and Walford (1951) and Ketchum, Redfield, and Ayers (1951). A concerted effect was also made in the early 1950s to understand and model the circulation and mixing within estuaries. This effort was in part stimulated by concern over the environmental impact of waste disposal within rivers and estuaries, and it produced a number of key ideas [e.g., Ketchum's (1950) tidal prism method to compute flushing times, and Stommel's (1953b) method for estimating the longitudinal diffusion coefficient in a well-mixed river or estuary from the observed salinity field]. Ketchum and Keen (1955) segmented the Middle Atlantic Bight from Cape Hatteras to Cape Cod and computed the flushing times for each segment, assuming only cross-shelf mixing and advection. They concluded that a considerable amount of cross-shelf transport of salt and river water must occur in both winter and summer to account for the observed mean salinity field. Ketchum and Corwin (1964) later examined the water structure south of Long Island over the period 1956–1959 and incorrectly concluded that the cool bottom water was formed only by local winter cooling, and then was warmed up by mixing with either warmer surface water or warmer slope water.

Both Bigelow and Iselin had long been aware of eddy-like features in the near-surface water structure over the shelf and slope region. In his interpretation of drift-bottle data obtained in the Middle Atlantic Bight in the spring of 1951, Miller (1952) suggested that a series of distinct current branches or eddies was superimposed on the general southwest alongshore drift. This work, plus the growing evidence of current variability in the Gulf Stream obtained by Fuglister and Worthington (1951) in Operation Cabot, led Iselin (1955) to urge that new observational methods be developed to study the Middle Atlantic Bight circulation. Iselin suggested a several-year program of continuous measurements of meteorological and oceanographic variables using new instruments deployed in moored arrays. Although others besides Iselin had also considered the potential of long-term moored-array measurement programs, the instrumentation and mooring technology required for such a program were simply not available yet. In 1954, D. Bumpus, C. Day, and J. Chase did start a cooperative program (Bumpus, 1955) with the U.S. Coast Guard to collect daily temperature and salinity measurements as well as meteorological observations at lightships and light stations in the Middle Atlantic Bight and Gulf of Maine. This collection program ran through the 1960s and provided the data used by Chase (1959), Howe (1962), Chase (1969), and Bumpus (1969) to examine the influence of runoff and wind on nearshore salinities and surface currents.

Bumpus, Miller, and others had frequently released drift bottles during hydrographic and other cruises in the Middle Atlantic Bight, and Bumpus and Lauzier (1965) summarized the results of both American and Canadian drift-bottle work conducted from 1948 to 1962. Frustrated by the fact that a standard drift bottle provides only a "birth-and-death" notice and little hard information about the drift in between, Bumpus (1956) and Bumpus et al. (1957) experimented with radio-tracked surface drifters (the "talking drift bottle") in the mid-1950s, but this effort was dropped by Bumpus as too expensive and inefficient (Bumpus, personal communication). Howe (1962) did use radio-tracked buoys with parachute drogues to make some short-term current measurements over the middle and outer shelf. In 1960, Bumpus launched a massive 10-year surface drift-bottle program (some 150,000 bottles released) over the eastern United States shelf, and in 1961 he began an equally massive 10-year bottom-drifter program (75,000 drifters released), using the newly developed Woodhead seabed drifter described by Lee, Bumpus, and Lauzier (1965). The essential idea behind this work was to seed the shelf water with a network of drifters at least monthly over a 10-year period in order to determine the annual cycle of surface and bottom drift from the inferred trajectories of the field of drifters. At about the same time, Lauzier began a separate, more modest long-term drifter program over the eastern Canadian shelf, and he and Bumpus decided to share their data collected in the Gulf of Maine region. The results of this work were presented in stages by Bumpus (1965, 1969), and in his final summary report (Bumpus, 1973).

Bumpus found that throughout the year there was a strong nearshore movement of bottom drifters into or at least toward the mouths of the major estuaries within the Middle Atlantic Bight. He also found that bottom drifters over the mid-shelf region in the Middle Atlantic Bight primarily moved southwestward with a mean speed of a few centimeters per second, with no significant seasonal variation in either pattern or inferred speed. Because very few bottom drifters deployed at depths greater than about 60 to 80 m over the outer shelf were recovered, Bumpus concluded that a line of divergence existed in the bottom flow. The recovery rate in the surface-drifter program was more meager because offshore winds in the fall through early spring dramatically reduced the percentage of drifters returned except from very near shore. The observed summer surface drift was southwest all along the coast except during prolonged periods of strong northward winds and low runoff, when the nearshore surface flow was reversed and became northeastward.

This drifter work summarized by Bumpus (1973) provides the first general picture of the mean and seasonal surface and bottom circulation in the Middle Atlantic Bight. Bumpus (1973) concluded that a mean longshore flow of order 5 $cm\,s^{-1}$ occurs between Cape Cod and Cape Hatteras, and that Nantucket Shoals and Cape Hatteras appear to be oceanographic barriers that limit in some sense the alongshore flow. Near Cape Hatteras the alongshore flow turns seaward and becomes entrained in the Gulf Stream, a process discussed first by Ford and Miller (1952) and more recently by Fisher (1972) and Kupferman and Garfield (1977).

7.3.3 Recent Developments

The drifter work undertaken by Bumpus and others in the 1960s provided the first quantitative description of the mean and seasonal surface and bottom circulation in the Middle Atlantic Bight. In the 1970s, moored arrays of self-contained current meters and other *in situ* instrumentation have been deployed as part of several new field programs, and these new *direct*-current measurements are providing the first detailed description of the regional circulation in the Middle Atlantic Bight. It is important to note that much of the new *in situ* instrumentation used in the 1970s evolved from development efforts begun in the 1950s and 1960s, and it was not until the late 1960s and early 1970s that U.S. oceanographers began to use instrumented moored arrays as a routine and reliable "tool" to measure current, temperature, and other physical variables in both the deep ocean and the shallower continental shelf and lakes.[6] The development of solid-state electronics, digital computers, and time-series analysis methods in the late 1950s and 1960s especially helped make the current-meter development efforts successful. This instrumentation revolution in the 1950s and 1960s has had a tremendous impact on the field of physical oceanography in the 1970s, for it has allowed in many cases a first direct look at a wide spectrum of oceanic motions and phenomena [see Gould (1976) and chapter 14].

The availability of these new observational tools plus an increased public concern over environmental issues helped motivate much of the new field work begun in the Middle Atlantic Bight in the 1970s. National concern over the environmental impact of marine waste disposal in the Middle Atlantic Bight became paramount in the late 1960s and early 1970s. The public media characterized the New York Bight—the coastal ocean between Long Island and New Jersey—as a "dead sea" caused by decades of sewage discharge and marine dumping of dredge spoils, building rubble, sewage sludge, industrial wastes, and other materials (Gross, Swanson, and Stanford, 1976). The NOAA Marine EcoSystems Analysis (MESA) Program was formed in 1972 to help focus both government and nongovernment research on regional problems caused by man's use of marine and estuarine resources, and, in 1973, the MESA New York Bight Project was started to develop a comprehensive research program to understand the New York Bight as a productive marine ecosystem. Other environmental concerns also helped initiate new field programs within the Middle Atlantic Bight. The New Jersey Public Service and Electric and Gas Company and the Long Island Electric Power Company both began to consider building floating nearshore nuclear-power plants, and New Jersey Public Service sponsored a 5-year field study of the physical oceanography off Little Egg Inlet, New Jersey, a potential power-plant site, while the Brookhaven National Laboratory began in 1973 a long-term program to study the nearshore currents, pollutant dispersion, and primary productivity off the southern coast of Long Island. The Bureau of Land Management of the Department of Interior also sponsored new field programs to identify and study the dominant sediment-transport processes in the Baltimore Canyon and Georges Bank regions in order to assess better the possible physical hazards and environmental impact of potential offshore petroleum development there.

These specific field programs plus others undertaken with National Science Foundation, Office of Naval Research, and other federal support have provided most of our new knowledge about the mean circulation and low-frequency current variability in the Middle Atlantic Bight. Using moored instrumentation, Boicourt (1973) and Beardsley and Butman (1974) showed that strong winter storms could produce alongshore currents of order 20 to 50 $cm\,s^{-1}$ in the mid-shelf region, and Boicourt and Hacker (1976) demonstrated that the low-frequency alongshore-current fluctuations were spatially coherent over a 200-km separation along the 38-m isobath in the southern Middle Atlantic Bight. Some preliminary results from several of these field programs were then presented by Beardsley, Boicourt, and Hansen (1976), who demonstrated that much of the subtidal current variability over the shallower portion of the Middle Atlantic Bight was directly wind driven in the synoptic (2- to-10-day) band. Their map of the directly observed mean subsurface currents demonstrated that at least during the 1-to-3-month duration of the initial moored-array experiments, the subsurface currents did flow alongshore toward the southwest, and that the average currents generally increased in magnitude offshore and decreased with closeness to the bottom. At most sites, the mean current veered toward shore with increasing depth. The summer current measurements made off New York and Cape Henry showed that the alongshore currents in the near-bottom cold pool equaled or exceeded the alongshore currents in the surrounding warmer water, clearly indicating that the summer cold pool is not the static feature originally envisaged by Bigelow and others.

Beardsley et al. (1976) also used the mean-current observations to estimate crudely the alongshore volume transport of shelf water through three transects across the Middle Atlantic Bight to the 100-m isobath. They found that the three estimates were surprisingly uniform considering the different time periods of the measurements and the different instrumentation used, and they suggested a value of $250 \times 10^3\ m^3\,s^{-1}$ for the average alongshore transport of shelf water within the 100-m isobath through the Middle Atlantic Bight. This mean value, when divided into the volume of the Middle Atlantic Bight, implied a mean residence time of the order of 0.75 years. This value, based on alongshore advection, is somewhat less than the value of 1.3 years based solely on cross-shelf mixing given by Ketchum and Keen (1955). Beardsley et al. (1976) also speculated that most of the shelf water observed flowing westward south of New England must, by continuity, flow into the Middle Atlantic Bight around Nantucket Shoals from the southern flank of Georges Bank and the Gulf of Maine region.

Longer current records have now been obtained in the Middle Atlantic Bight, and, in the rest of this section, we shall reexamine the preliminary circulation picture described in 1976. Since much of the subtidal current variability over the shelf is wind driven, we shall first describe the nature and structure of the atmospheric forcing found over the Middle Atlantic Bight in some detail, and then describe what is known about the wind-driven shelf response on three time scales: the synoptic 2-to-10-day time scale, the monthly-mean time scale, and the long-term mean time scale.

Atmospheric Forcing The temporal and spatial structure of the surface-wind-stress and pressure fields found over the Middle Atlantic Bight will be described next. Atmospheric motions are commonly classified into micro-, meso-, or synoptic-scale motions, depending on their characteristic time and space scales. While the exact limits of these scales are somewhat ill defined (Fiedler and Panofsky, 1970; Orlanski, 1975), synoptic-scale motions are generally characterized by periods in excess of 2 days and horizontal-length scales in excess of 500 km. The synoptic-scale weather disturbances are generally caused by baroclinic instability, while the principal mesoscale phenomena in the Middle Atlantic Bight—the summer seabreeze and cellular convection—are primarily associated with mechanical and hydrostatic instabilities (Mooers, Fernandez-Partagas, and Price, 1976; Mayer, Hansen, and Ortman, 1979). Because the early direct oceanic wind measurements by Millard (1971) and more recent work demonstrate that the synoptic-scale disturbances cause most of the observed surface-wind variance over the shelf and open ocean, we shall focus this discussion on the synoptic-scale surface weather over the Middle Atlantic Bight.

Synoptic-scale surface forcing: Mooers, Fernandez-Partagas, and Price (1976) have examined in great detail a variety of atmospheric-data sets to construct the first comprehensive picture of synoptic-scale atmospheric forcing over the Middle Atlantic Bight. The Atlantic coastal region between North Carolina and New England is well known for intense cyclogenesis. The combination of cool continental air over the eastern United States and warm moist maritime air offshore causes the synoptic-scale low-pressure disturbances or cyclones both to intensify with time and to tend to propagate toward the northeast along the coast.' The temperature contrast between continental and maritime air masses is greatest in winter, so that winter cyclones are usually the most intense and frequent (averaging 5 per month), and cause most of the synoptic-scale variability. The summer cyclones are generally much weaker and less frequent (averaging 2.5 per month), though strong storms can occasionally occur in summer (see Mather, Adams, and Yoshioka, 1964). Hurricane-strength winds occur in the Middle Atlantic Bight on average once in 6 years.

Mooers et al. (1976) have constructed a model winter cyclone based on a case study of 34 synoptic disturbances occurring in the Middle Atlantic Bight during the winters of 1972–1974. While quite simplified, their model cyclone exhibits a number of features that appear to characterize many winter cyclones. The model cyclone forms near Cape Hatteras and propagates toward the east and north as shown in figure 7.6. The central pressure deepens, and both the spatial scale of the cyclone and the strenth of the surface wind stress increase with time. The surface wind stress and the surface heat flux into the atmosphere are a maximum in the western sector of the storm, where cold continental air flows out across the shelf and pronounced mesoscale cellular convection occurs (Burt and Agee, 1977). Pronounced warm-front-type precipitation occurs in the northern sector and causes the net precipitation to exceed evaporation along the storm track. The model cyclone develops a surface wind-stress pattern which is coherent in both time and space over the extent of the Middle Atlantic Bight, and the maximum wind stresses are larger than the winter mean stress in the Middle Atlantic Bight by a factor of 2 to 4. While real storms are more complex and follow a variety of different paths both north and south of the Middle Atlantic Bight, the model cyclone provides a useful conceptual framework of the spatial and temporal evolution of the surface wind stress and wind-stress curl fields from an event point of view.

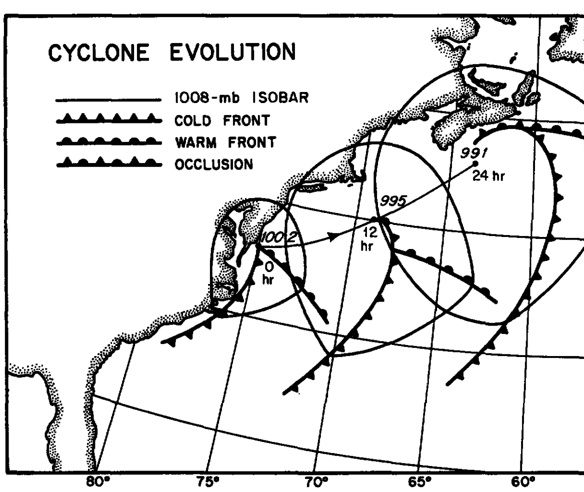

Figure 7.6 The evolution of the model winter cyclone developed by Mooers, Fernandez-Partagas, and Price (1976). The 1008-mb contour outlining the cyclone is shown at 0, 12, and 24 hours after the storm center has left the coast. The deepening of the central pressure is also shown along the storm track.

Mooers et al. (1976) also examined surface pressure and wind data collected at coastal stations along the Middle Atlantic Bight and the Gulf of Maine region, and found, using coherence and correlation analysis, that (a) the surface atmospheric pressure (p_a) field was spatially coherent over the entire Middle Atlantic Bight for time scales greater than 2 days, (b) the p_a and east and north wind-stress components τ_E and τ_N have (zero-time lag) correlation scales of about 1600, 600, and 800 km, respectively, and (c) the p_a-, τ_E-, and τ_N-fields generally propagate toward the northeast with speeds of 15, 5, and 10 $m\,s^{-1}$. They argue that the stress field is not "frozen" into the cyclone as defined by the surface pressure field, and that the difference in phase speeds between p_a and the stress components is associated with the tendency for winter storms to intensify in time, as illustrated in the model cyclone.

Two NOAA environmental buoys designated EB34 and EB41 were deployed in the central Middle Atlantic Bight in late 1975. EB34 (40.1°N, 73.0°W) was located 50 km southeast of New York City and EB41 (38.7°N, 73.6°W) was located 100 km off the New Jersey coast (see figure 7.7). Both buoys were equipped with vortex-shedding anemometers at a height of 5 m, and routinely transmitted 8.5-minute averages of wind speed and direction to shore. The initial wind data collected with these buoys have been studied by Mooers et al. (1976), Williams and Godshell (1977), Overland and Gemmill (1977), Noble and Butman (1979), and Mayer et al. (1979); Wang (1979c) has examined wind variability at the Chesapeake Lighttower. These studies collectively demonstrate three key features of the synoptic-scale surface pressure and wind-stress fields: (a) the winter-buoy data are highly coherent with coastal data, a result which is consistent with the correlation-scale information obtained by Mooers et al. (1976), and implies that their results apply to the surface pressure and wind-stress fields over the shelf; (b) the synoptic-scale surface wind stress increases in magnitude by a factor of 2 to 3 across the shelf; and (c) a significant cyclonic verring of the surface stress field by as much as 30° may occur across the shelf.

We show in the upper half of figure 7.8 wind-stress spectra obtained at Atlantic City, EB41, and at Ocean Weathership C (52.5°N, 35.5°W) located east of the Grand Banks in the western North Atlantic. The wind

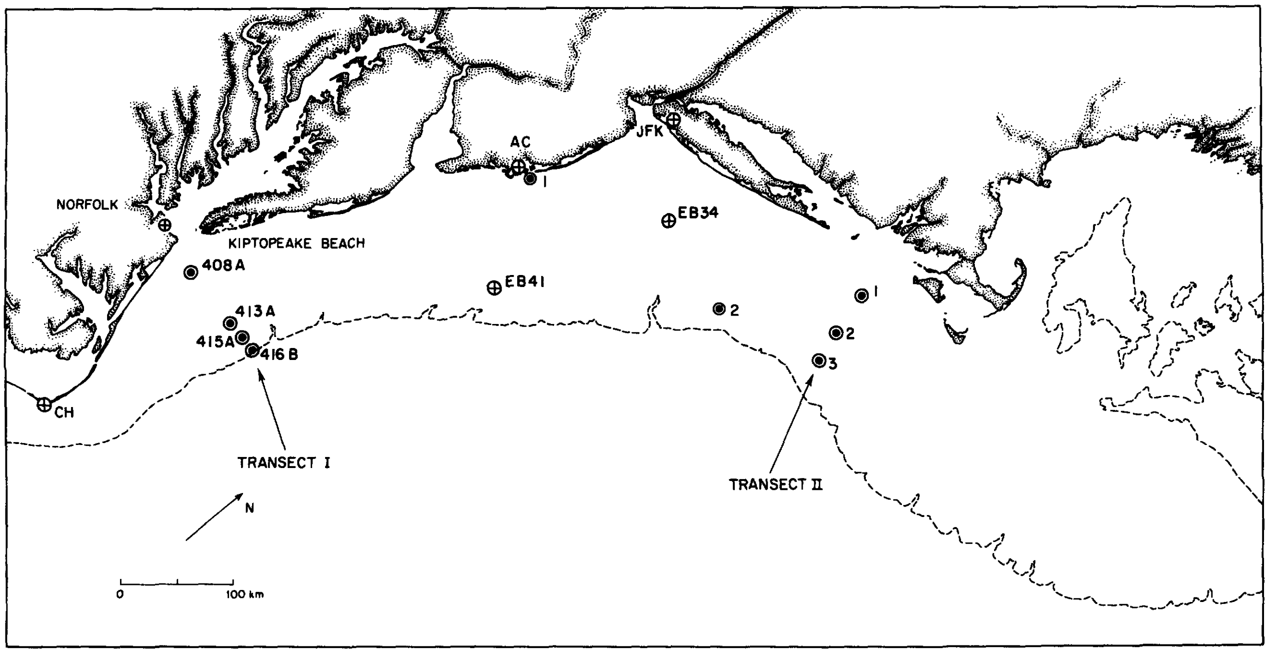

Figure 7.7 Map of the Middle Atlantic Bight showing the 200-m depth contour (dashed) and the wind (⊕), sea-level (▲), and current-meter (◉) stations discussed in the text.

stress has been computed using the quadratic drag law, the observed or adjusted 10-m-high wind vector, and a constant drag coefficient of 1.5×10^{-3}. The Atlantic City and EB41 spectra are computed by Noble and Butman (1979) from 6-month-long time series obtained from December 1975 through May 1976, while the Weathership C spectrum reported by Willebrand (1978) is based on a 25-year-long time series. See table 7.1 for more details. We have included the Weathership C spectrum since the weathership was located some 2500 km northeast of the Middle Atlantic Bight along the principal storm track followed by winter cyclones that develop along the Atlantic coastal region, and it represents the closest station (known to us) for which a well-resolved wind-stress spectrum has been published. The spectra have been smoothed within the estimated statistical uncertainty to simplify the graphical presentation. (See the additional spectra in chapter 11.)

The three spectra demonstrate that most of the wind-stress variability is caused by rather broadband synoptic-scale atmospheric transients with characteristic periods between 2 and 10 days. The spectra are red (with increasing power densities at decreasing frequencies), and show an approximate -2 power-frequency decay at frequencies above about 0.5 cpd. The two winter shelf spectra do not exhibit a diurnal peak since the sea breeze is a summer phenomenon (Mayer et al., 1979). While the Atlantic City and EB41 time series are too short to determine clearly the very low-frequency dependence, the Weathership C spectrum follows a well-defined -0.4 power-frequency dependence between 0.2 cpd and the annual frequency 0.003 cpd.

The two Middle Atlantic Bight spectra demonstrate the marked increase in wind-stress magnitude from nearshore toward the shelf break. The EB41 power density is about a factor of 8 larger than the Atlantic City density over most of the frequency bands resolved. The Weathership C power density is even a factor of 5 larger than the EB41 power density. Willebrand (1978) has examined surface pressure and wind-stress data obtained at both Weathership C and Ocean Weathership D (44°N, 41°W), located roughly about half-way between the Middle Atlantic Bight and Weathership C. He finds that synoptic-scale disturbances in the 2-to-10-day band do propagate toward the northeast in the western North Atlantic but that at periods greater than about 10 days, atmospheric pressure fluctuations seem to have no preference for east-west phase propagation. Between the two weatherships, τ_N was coherent at all periods greater than 1 day, while τ_E was incoherent at periods greater than 10 days. These results suggest that the larger synoptic-scale transients continue to intensify as they move northeastward toward the Grand Banks region.

We have focused on the traveling winter cyclone as the predominant synoptic-scale disturbance that influences the Middle Atlantic Bight, and have described the frequency structure and correlation spatial-scale information presently known in some detail. The surface pressure and wind-stress-correlation space scales are sufficiently large in comparison to the cross-shelf and alongshelf dimensions of the Middle Atlantic Bight

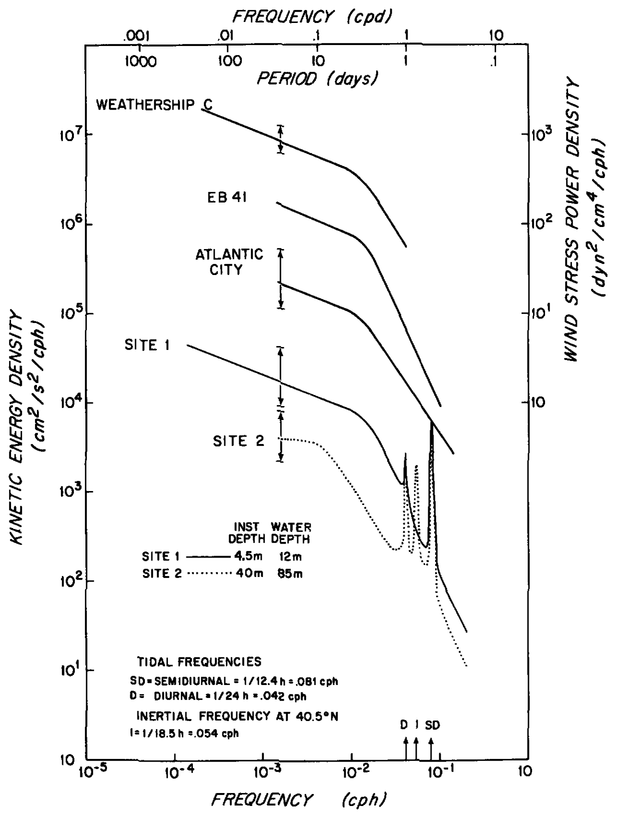

Figure 7.8 Wind-stress and current spectra. The three upper curves represent the wind-stress spectra for Ocean Weathership C, environmental buoy EB41, and Atlantic City, New Jersey. The two lower curves are the kinetic-energy spectra obtained at a nearshore site (site 1) off New Jersey and a deeper site (site 2) located near the shelf break south of New England (see table 7.1). The vertical brackets indicate the 95% confidence limits.

that the synoptic-scale surface forcing should be spatially coherent over much of the shelf.[7] Frankignoul and Müller (1979) have recently reviewed the meager information available on the wavenumber structure of the surface wind-stress field over the ocean, and suggest several forms for the wavenumber spectra for the synoptic-scale surface pressure, wind-stress, and wind-stress-curl fields. Their model wind-stress spectrum is essentially white for wavenumbers $k \ll k_b = 2\pi/5000\ \mathrm{km}^{-1}$ (which is a wavenumber magnitude characteristic of mid-latitude baroclinic instability), and decays at smaller scales like k^{-2} for $k > k_b$. The smaller-scale surface wind-stress fluctuations appear to be horizontally isotropic at higher wavenumbers $k > 2\pi/200\ \mathrm{km}^{-1}$. If we assume that the shape of the frequency spectrum obtained at Weathership C and the model wavenumber spectrum suggested by Frankignoul and Müller (1979) are both applicable to the Middle Atlantic Bight region, then we find that about 50% of the total wind-stress variance in the 1-day-to-1-year band is caused by synoptic-scale transients concentrated in the 2-to-10-day band, and some 70% of the wind-stress variance in the $2\pi/200$-to-$2\pi/10{,}000\ \mathrm{km}^{-1}$ wavenumber band is caused by atmospheric motions with scales larger than 1600 km, which is twice the alongshelf length of the Middle Atlantic Bight.

Seasonal and mean surface forcing: Saunders (1977) has computed the seasonal surface wind-stress pattern with a 1°-square resolution over the eastern continental shelf of North America using 32 years of ship wind reports. His annual mean and three-monthly mean wind-stress maps show that, except in summer, the mean and seasonal wind stresses in the Middle Atlantic Bight are generally to the east and southeast and exhibit a significant increase in magnitude and some cyclonic veering with increasing distance from the coast. The relatively weak summer wind stress is directed toward the northeast and exhibits some anticyclonic veering relative to the coast. The standard deviation of the stress in nearshore and offshore regions is 1.5 and 2.5 $\mathrm{dyn\,cm^{-2}}$, respectively, in winter, and 0.5 and 1.5 $\mathrm{dyn\,cm^{-2}}$ in summer. The offshore increase in the stress is most obvious in winter and spring seasons, and, since these periods dominate the mean stress, they cause the pronounced offshore increase observed in the mean stress pattern. The mean and seasonal wind-stress pattern over the adjacent western North Atlantic is described by Leetmaa and Bunker (1978) and will be discussed below. The annual air–sea interaction cycles and continental runoff are described in the appendix for completeness.

The Synoptic-Scale Shelf Circulation We shall now focus on the response of the Middle Atlantic Bight to synoptic-scale (2-to-10-day) atmospheric forcing. We shall first examine what is known about the temporal and spatial structure of the transient wind-driven shelf circulation, and then describe a conceptual model suggested by the existing current and sea-level data.

Temporal structure: We show in the bottom half of figure 7.8 kinetic-energy spectra computed from current records 6 months or longer obtained at two representative sites in the Middle Atlantic Bight. The two sites are labeled 1 and 2 and the locations and other pertinent information for each site are given in table 7.1. Site 1 is located 4.5 km off Little Egg Inlet, New Jersey, and site 2 is located on the outer New England shelf about 50 km east of the Hudson Canyon (see figure 7.7). The kinetic-energy spectra for sites 1 and 2 are from EG&G (1978) and Ou (1979) respectively, and have been smoothed within the estimated uncertainties to simplify the graphical presentation. It is important to remember that a significant spectral gap exists between energetic high-frequency motions characterized by periods from several seconds to minutes (associated with surface and internal gravity waves, and related wave and turbulent phenomena) and energetic lower-frequency motions characterized by periods

Table 7.1 Location and Other Pertinent Information for the Three Wind-Stress and Two Current-Kinetic-Energy Spectra Shown in Figure 7.8

Site	Location	Time span	Inst. height-depth (m)	Water depth (m)	Data source
Wind stress					
Atlantic City	39°27′N, 74°34′W	Dec. 75–May 76	6	—	Noble & Butman (1979)
EB41	38°42′N, 73°36′W	Dec. 75–May 76	5	—	Noble & Butman (1979)
Weathership C	52°30′N, 35°30′W	Jan. 45–Dec. 71	10	—	Willebrand (1978)
Current					
1	39°28′N, 74°15′W	Apr. 73–Dec. 76	4.5	12	EG&G (1978)
2	39°59′N, 71°54′W	Feb. 76–Aug. 76	38	83	Ou (1979)

greater than several hours. We do not show here the higher-frequency end of the kinetic-energy spectra, but simply note that significant kinetic energy can exist at high frequencies, primarily associated with gravity-wave phenomena. While the importance of surface waves on beach erosion and nearshore sediment transport is generally appreciated (Lavelle et al., 1976), the recent field observations by Lavelle, Young, Swift, and Clarke (1978) and Butman, Noble, and Folger (1979) demonstrate sediment resuspension by storm-generated surface waves in depths out to 85 m in the Middle Atlantic Bight, and Grant and Madsen (1979) describe how the combined motion of waves and a lower-frequency current over a rough bottom can lead to an increased bottom drag on the current. While the existence of the spectral gap allows the high-frequency current signal to be removed by vector averaging, the 5-to-15-second oscillatory currents and the mooring motion associated with surface gravity waves are the primary source of contamination in the measurement of lower-frequency currents with existing mechanical current meters and standard mooring techniques.

The two kinetic-energy spectra shown in figure 7.8 illustrate several fundamental features of the transient-current variability observed in the Middle Atlantic Bight. Both spectra are inherently red, with the kinetic-energy density generally increasing with decreasing frequency below 0.5 cpd. The marked similarity in shape between the three wind-stress spectra and the kinetic-energy spectrum at site 1 (especially the −0.4 frequency dependence below the break near 0.3 cpd) suggests that the subtidal current fluctuations over the shallow inner shelf are directly wind driven over a very wide frequency band, from 0.01 to 1.0 cpd. Approximately 50% of the subtidal current variance observed at site 1 occurs in the synoptic-scale (2-to-10-day) band. At site 2 near the shelf break, the subtidal kinetic-energy density is smaller because of the increased water depth and measurement level, and the spectral shape is different in that the transition or break in the subtidal portion of the spectrum occurs at a lower frequency, near 0.1 cpd. This indicates that the subtidal current variability observed over the outer shelf is caused by both direct wind forcing and by the transmission or leakage of lower-frequency motion onto the outer shelf from the deeper ocean. The propagation of topographic Rossby waves up the continental rise and slope, the meandering of the Gulf Stream, and the passage of anticyclonic warm-core eddies near the shelf break can conceivably generate strong low-frequency currents over the outer shelf. These three different mechanisms and some supporting observations are described by P. C. Smith (1978), Ou (1979), Halliwell (1978), and Scarlet and Flagg (1979).

Flagg (1977), EG&G (1978), Bennett and Magnell (1979), Mayer et al. (1979), Butman et al. (1979), and Chuang, Wang, and Boicourt (1979) have examined the coherence between the local wind stress and the alongshelf and cross-shelf current components and all find that alongshelf currents at all observed levels are significantly coherent with the local alongshelf wind stress over the synoptic-scale band. The reported coherence squared generally varies from about 0.5 to 0.8, with a poorly defined time lag of 4 to 10 hours between the alongshelf wind-stress and current components. To a lesser extent the cross-shelf current component is also significantly coherent with the local alongshelf wind-stress component. Neither current component is significantly coherent with the local cross-shelf wind stress except perhaps very near the surface (within a few meters) and near the coast in the New York Bight (EG&G, 1978; Csanady, 1980), where cross-shelf winds can set up trapped pressure fields that drive alongshore currents.

To illustrate the strong coherence between alongshore wind and currents, we show in figure 7.9 the coherence and phase between the alongshelf wind or wind stress and the alongshelf current observed at the two shelf sites just discussed. The coherences shown are significant and relatively high over the synoptic (2-to-10-day) band at both sites, and the phases indicate that the alongshelf currents lag the alongshelf wind or wind stress by about 4 hours at site 1 and about 10 hours at site 2. While the coherence at site 1 exhibits some seasonal change [with a tendency for higher coherence in the synoptic-scale band during less stratified winter periods (EG&G, 1978)], the mean coherence squared is relatively constant over a wide frequency band between 0.03 and 0.5 cpd, indicating the alongshore currents just off New Jersey are directly wind driven at very low frequencies down to the monthly time scale. The current record at site 2 is too short to resolve accurately the lower-frequency coherence cutoff.

Spatial structure: We have pictured next in figures 7.10 and 7.11 composite vector diagrams to illustrate the vertical and cross-shelf structure of the synoptic-scale current fluctuations in the Middle Atlantic Bight. In figure 7.10 the adjusted sea level at Kiptopeake Beach near Cape Charles, the Norfolk wind stress, and the subtidal currents observed on a cross-shelf transect off Cape Henry are shown for the summer period 24 July to 22 August 1974. In figure 7.11 the local wind stress (computed from surface pressure charts) and the subtidal currents observed on a cross-shelf transect south of New England are shown for the winter period 1 March to 31 March 1974. The locations of the two transects are shown in figure 7.7. The mooring number and local water depth are given on the right of each figure and the depth of the current measurements is given on the left. The local alongshelf direction is vertical in both figures and the current time series have been low-pass filtered to remove the tidal and higher-frequency components above about 0.7 cpd. The means have not been removed from each time series.

These two vector diagrams illustrate the following common features of the synoptic-scale current fluctuations. There is a clear coherence between synoptic-scale wind-stress events and current fluctuations, with both strong up- and downshelf currents being driven by up- and downshelf wind stresses.[8] The alongshelf current fluctuations are themselves visually coherent and roughly in phase in both the vertical and cross-shelf directions. The amplitudes of the alongshelf current fluctuations do not vary significantly in the cross-shelf plane except within the bottom boundary layer, and the empirical orthogonal-function computations of Flagg (1977) and Chuang et al. (1979) indicate that a single barotropic vertical mode can account for approximately 90% or more of the subtidal alongshelf-current variance in both winter and summer. The observed cross-shelf currents have a more complicated vertical structure, with strong upshelf wind stress driving offshelf flow near the surface and onshelf flow near the bottom, and strong downshelf wind stress driving onshelf flow near the surface and offshelf flow near the bottom. This tendency for the profile of the cross-shelf-current fluctuation to reverse with depth has been noted by Boicourt and Hacker (1976), Scott and Csanady (1976), Flagg (1977), EG&G (1978), Mayer et al. (1979), and Chuang et al. (1979). Flagg (1977) and Chuang et al. (1979) find that the lowest baroclinic empirical orthogonal mode and the barotropic mode must both be used to account for more than 90% of the cross-shelf-current variance. Mayer et al. (1979) and Chuang et al. (1979) find that the depth of the cross-shelf-velocity node varies with stratification over

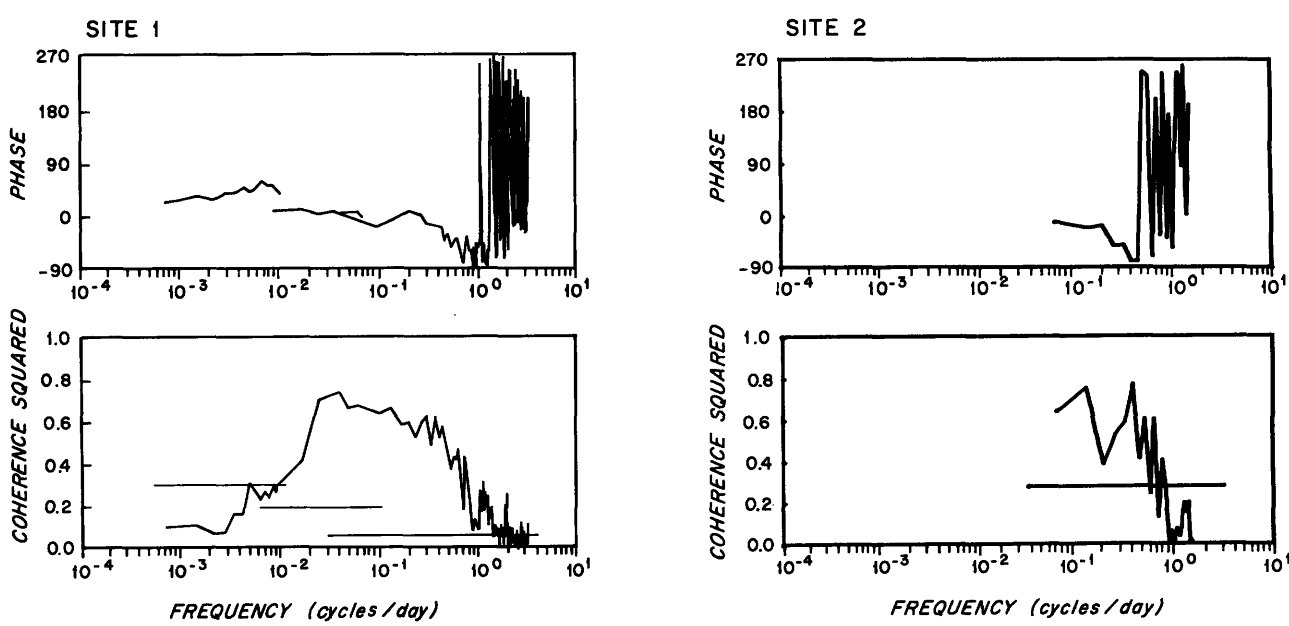

Figure 7.9 Coherence and phase computed between local alongshelf wind and current components at site 1 and between local alongshelf wind-stress and current components at site 2. Locations of the two shelf sites are given in Table 7.1 and the corresponding kinetic energy spectra at both sites are shown in figure 7.8. The horizontal lines indicate the 95% confidence limit for zero true coherence, and a negative phase indicates that the current lags the wind or wind-stress.

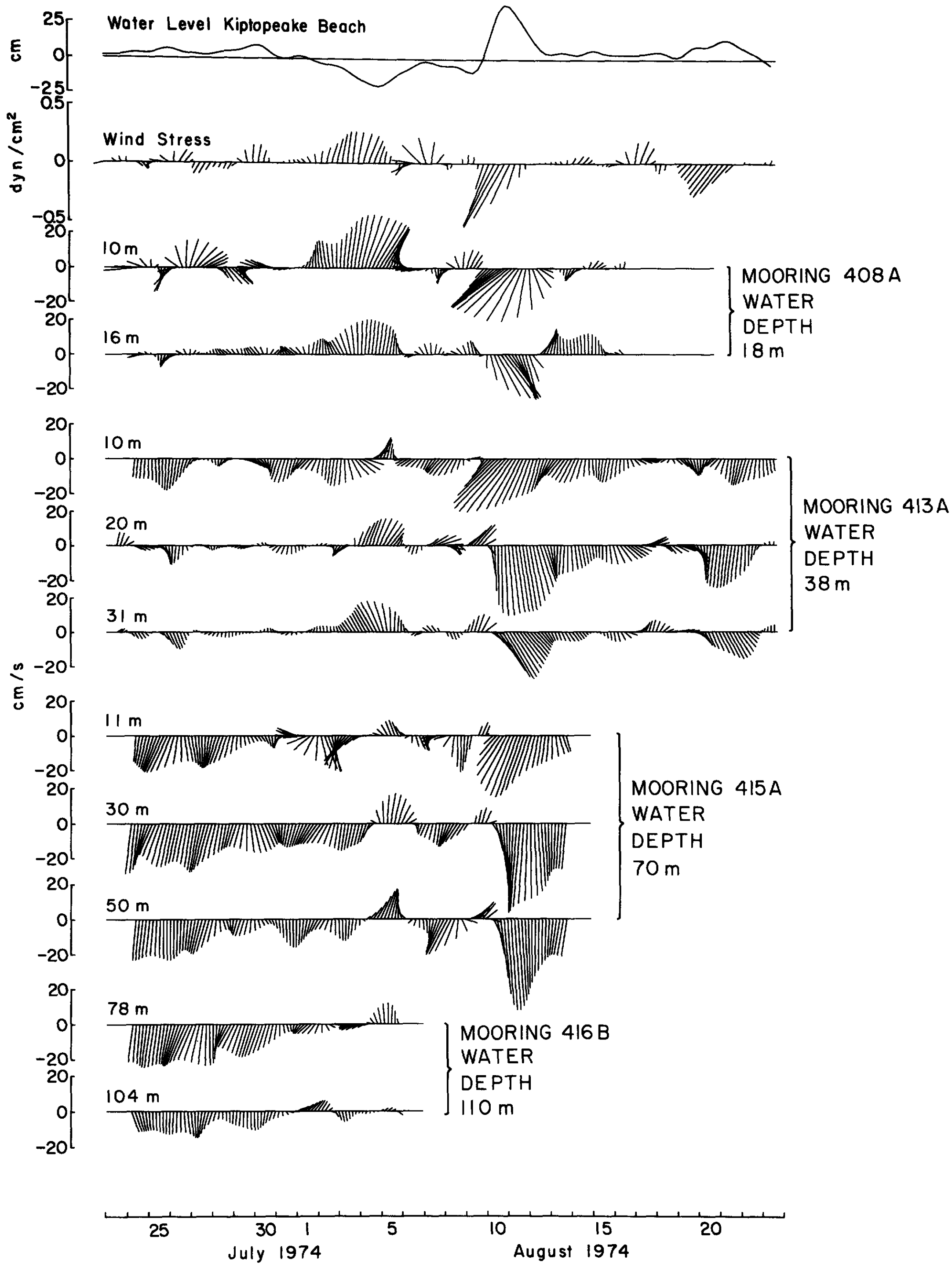

Figure 7.10 Summer vector time series of adjusted sea level at Kiptopeake Beach, Norfolk wind stress, and subtidal currents measured along the cross-shelf transect I located off Cape Henry. Current-measurement depths shown to the left and mooring numbers and local-water depths shown to the right. The transect mooring locations are shown in figure 7.7. North is upward, approximately parallel with the local alongshelf direction.

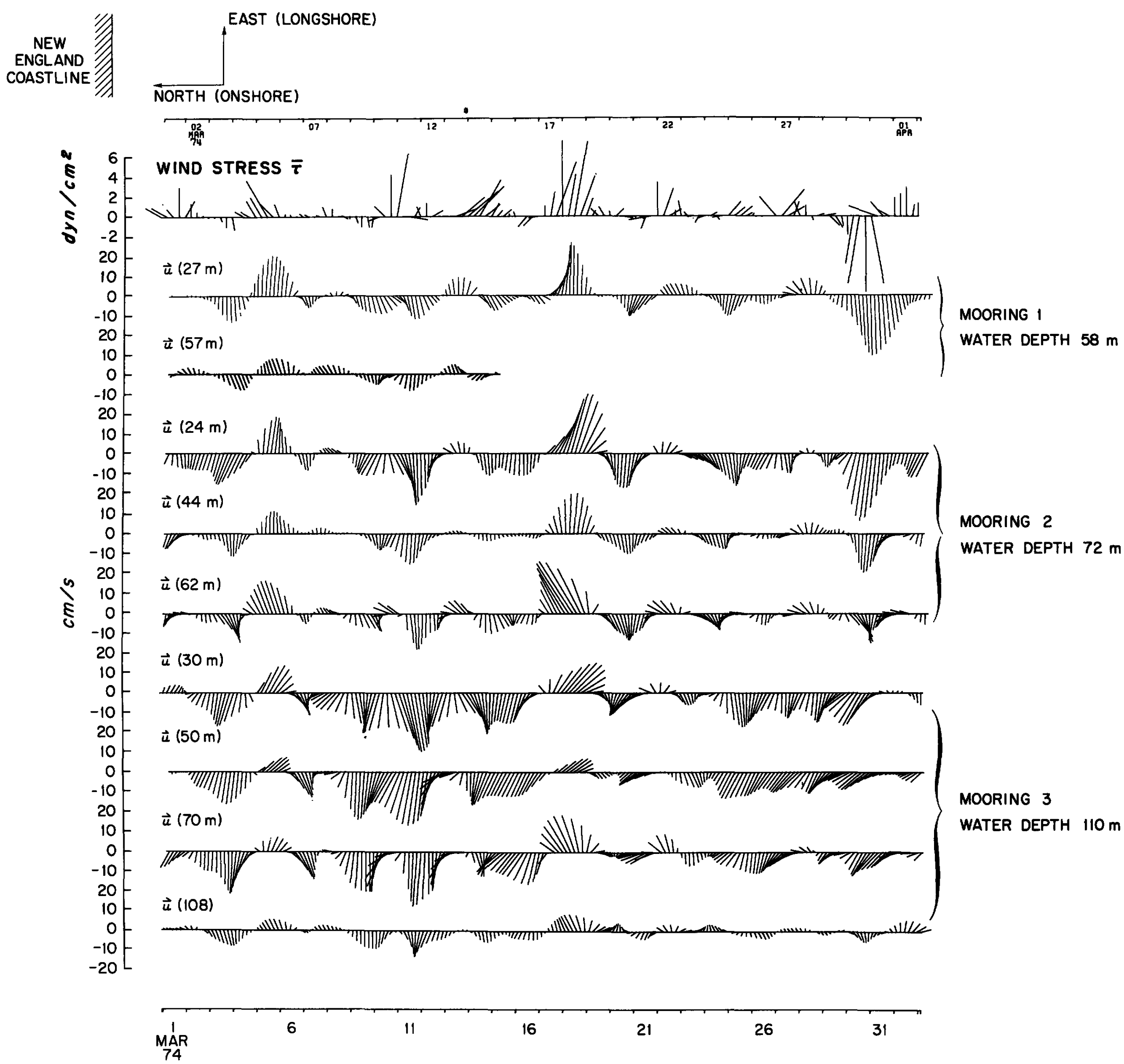

Figure 7.11 Winter vector time series of local wind-stress and subtidal currents measured along the cross-shelf transect II located approximately 100 km west of Nantucket Shoals. Current-measurement depths shown to left and mooring numbers and local-water depths shown to right. The transect and the mooring locations are shown in figure 7.7. East is oriented upward, approximately parallel with the local alongshelf direction.

the mid-shelf. The node is relatively deep during winter, and is significantly shallower in the summer when the seasonal pycnocline is well established.

Much less is directly known about the alongshelf structure of the synoptic-scale current fluctuations within the Middle Atlantic Bight. Boicourt and Hacker (1976), Butman et al. (1979), and Ou (personal communication) find that along an isobath over the middle and outer shelf, the alongshelf current component is highly coherent over alongshore separations up to 235 km, the largest separation examined, while the cross-shelf current component is incoherent over a 70-km separation, the smallest separation examined.

In summary, the synoptic-scale alongshelf-current fluctuations are generally coherent with the local alongshelf wind stress, and appear to have a relatively simple spatial structure throughout the year. The alongshelf-current fluctuations are essentially barotropic and spatially coherent in the cross-shelf plane. The alongshelf currents are also coherent over alongshelf separations up to 235 km, although the structure of the synoptic-scale atmospheric forcing suggests that significant coherence should be found over much larger separations. The generally weaker cross-shelf current component, although spatially coherent in the cross-shelf plane, appears to have a much smaller alongshelf coherence, of order 50 km or less.

A conceptual model: These current observations can be interpreted within the conceptual framework provided by continental shelf-wave theory. It is now recognized from a number of theoretical and experimental studies [see reviews by Mysak (1980) and Allen (1980) and chapters 10 and 11] that continental margins can act as effective waveguides for the alongshelf propagation of subinertial current fluctuations. The sloping topography of the continental margin and the density stratification over the shelf and slope are two basic conditions that lead to coastally trapped wave motions. The offshore increase in depth can support barotropic vorticity waves known as continental shelf waves in a homogeneous fluid, while internal Kelvin waves can propagate along a vertical boundary in a stratified fluid. Since the internal Rossby radius of deformation over the inner shelf in the Middle Atlantic Bight is of order 10 km or less, and is thus considerably smaller than the width of the shelf, the internal Kelvin wave activity, if present, should be trapped in a relatively thin coastal boundary layer. The theoretical and numerical calculations made by A. J. Clarke (1977) and Wang and Mooers (1977) imply that even with realistic stratification, the alongshelf currents for both forced and free continental shelf waves in the Middle Atlantic Bight should be essentially barotropic over the shelf, and coherent with the local coastal sea level.[9] The cross-shelf momentum balance is essentially geostrophic in both free and forced shelf waves, so that in theory the subsurface pressure fluctuations should have a simple monotonic cross-shelf structure, with a maximum amplitude at the coast and a vanishing amplitude off the shelf. Beardsley et al. (1977) have examined the spatial structure of the synoptic-scale subsurface pressure fluctuations observed over the northern half of the Middle Atlantic Bight, and they found that (a) the subsurface pressure fluctuations observed over the shelf were coherent and in phase with coastal sea level, and (b) the subsurface pressure fluctuations had a monotonic cross-shelf structure, with very small amplitudes observed near the shelf break. This last result has been further substantiated with bottom pressure measurements made by Brown (personal communication) over the outer New England shelf and upper slope. Flagg (1977) found significant coherence (coherence squared $\simeq 0.7$–0.8) between local subsurface pressure and alongshelf-current fluctuations at a mid-shelf site on the New England shelf, and Chuang et al. (1979) also found high coherence (coherence squared $\simeq 0.7$–0.8) between synoptic-scale alongshelf currents and coastal sea level off Cape Henry.

These direct-current and pressure observations are consistent with the conceptual model of coastally trapped continental shelf waves. In view of the clear relation observed between the essentially barotropic alongshelf-current and coastal-sea-level fluctuations, the coastal-sea-level studies of Wang (1979c) and Noble and Butman (1979) can be used to infer the alongshelf structure of the synoptic-scale transient shelf circulation over larger spatial scales than have been examined yet through direct-current measurements. Wang (1979c) and Noble and Butman (1979) have investigated the relation between local wind stress and sea level in the Middle Atlantic Bight; they find that north of Cape May coastal sea level and the alongshelf wind stress are highly coherent over the synoptic-scale band, and coastal sea level lags the local alongshelf wind stress by 8 to 12 hours, indicating that the alongshelf-current and coastal-sea-level fluctuations are in phase to within a few hours. Both coastal wind stress and coastal sea level move slowly upshelf toward the northeast, and the very slow spatial decay in coastal-sea-level coherence suggests that the synoptic-scale alongshelf-current fluctuations are spatially coherent over the entire northern section of the Middle Atlantic Bight.

The observed coastal-sea-level response south of Cape May is more complex. Wang (1979c) finds that local alongshore-wind-stress and coastal-sea-level coherence is only high at frequencies above 0.3 cpd, while the lower-frequency coastal-sea-level fluctuations appear to propagate downshelf toward the south (like free shelf waves) with a phase speed of order 600 km day^{-1}. Downshelf propagation of the band-averaged (0.04 to 0.4 cpd) coastal-sea-level fluctuations is also reported by Noble and Butman (1979).

These results suggest the following physical interpretation. Free continental shelf waves have a downshelf or southward phase velocity in the Middle Atlantic Bight, while forced shelf waves driven by propagating atmospheric disturbances tend to move in either alongshelf direction, in phase with the forcing. The nature of the forced shelf-wave response depends critically on several factors: the shelf topography and coastline geometry, the spatial and temporal structure and propagation characteristics of the wind-stress pattern, and the effective frictional-adjustment time scale of the shelf. In the limit of no friction, both free and forced shelf waves are generated by wind-stress patterns moving along the shelf, and the alongshelf current generally lags the local wind stress by 90° (Gill and Schumann, 1974). In the limit of steady forcing and/or rapid frictional adjustment only the forced wave exists [called an *arrested topographic wave* in this limit by Csanady (1978)], and the phase lag between the alongshelf current and wind-stress components is significantly reduced (Gill and Schumann, 1974; Hsueh and Peng, 1978; Brink and Allen, 1978). The frictional time scale for the shelf north of Cape May appears to be sufficiently short that a quasi-steady-state response to synoptic-scale atmospheric forcing is observed. The analytic and numerical-model calculations of Csanady (1974) and Beardsley and Haidvogel (1980) suggest that the frictional-adjustment time scale for much of the Middle Atlantic Bight is roughly $T_f = 10$ hours. Thus the ratio of T_f to the characteristic time scale T_a of the forcing ($T_a = L/c$ where L and c are the alongshelf length scale and phase speed of the forcing) is small enough (≤ 0.2) for most storms that the wind-driven response has much of the character of a heavily damped or arrested shelf wave, slowly moving along the shelf in phase with the forcing. The shelf south of Cape May appears to exhibit both this quasi-steady-state response and a southward-moving shelf-wave response at lower frequencies due to upshelf generation. Why this occurs over the southern region of the Middle Atlantic Bight is not yet clear, although it must be related to the three-dimensional character of the shelf topography north of Cape Hatteras, as well as to the spatial structure of the atmospheric forcing. The scattering of kinetic energy from longer to shorter shelf waves by topographic perturbations like canyons and capes may explain the small alongshelf coherence scale for the cross-shelf current field (Wang, 1980).

We have focused in this section on the large-scale response of the Middle Atlantic Bight to synoptic-scale atmospheric forcing and have attempted to interpret recent current and pressure observations in light of continental shelf-wave theory (with its implicit three-dimensionality). It seems clear that more observational and theoretical work is needed before the synoptic-scale response is thoroughly understood. It does seem feasible, however, that the alongshelf current fluctuations within the Middle Atlantic Bight may be predictable from the atmospheric forcing and coastal sea level once the relative roles of free and forced shelf waves and their generation and dissipation mechanisms in this particular shelf domain are clarified. Coastally trapped pressure fields and the transient circulations associated with them also occur on shorter space and time scales. These fields are generally created by either sharp spatial or temporal variations in the wind-stress pattern or apparent spatial changes in the coastal wind-stress pattern due to changes in the coastline orientation. The almost 90° change in the coastline orientation at New York causes a smaller-scale trapped pressure and current field to exist within the New York Bight (Wang, 1979c; Csanady, 1980).

The Monthly-Mean Circulation We focus in this section on the monthly-mean current field in the Middle Atlantic Bight. We have chosen to use one month as a convenient averaging period for the following reason. In the previous section, we noted that the local alongshore wind-stress and current components over much of the shelf are coherent and nearly in phase over the synoptic-scale frequency band from 0.1 to 0.5 cpd. At lower frequencies, the alongshore wind-stress and current components generally become incoherent although off New Jersey the nearshore wind stress and currents remain significantly coherent at lower frequencies down to at least 0.03 cpd (a period of 1 month). From this we should expect to see some coherence between the monthly-mean wind stress and currents over the middle and inner shelf and we should be able, perhaps, to attribute some of the observed monthly-mean current variability to the monthly-mean wind-stress fluctuations.

Mayer et al. (1979) have discussed the monthly-mean current variability observed at the MESA long-term mooring site in the New York Bight. We show here in figure 7.12 a composite vector diagram incorporating both the data given by Mayer et al. (1979) and other current data obtained during the same time span at a long-term site located off Cape Henry by Boicourt and Chuang (personal communication). The additional wind-stress data shown for the two environmental buoys and Cape Hatteras have been supplied by Halliwell (personal communication) and Noble and Butman (personal communication). The wind-stress and current vectors at the different measurement levels are plotted with true north directed upward, and the

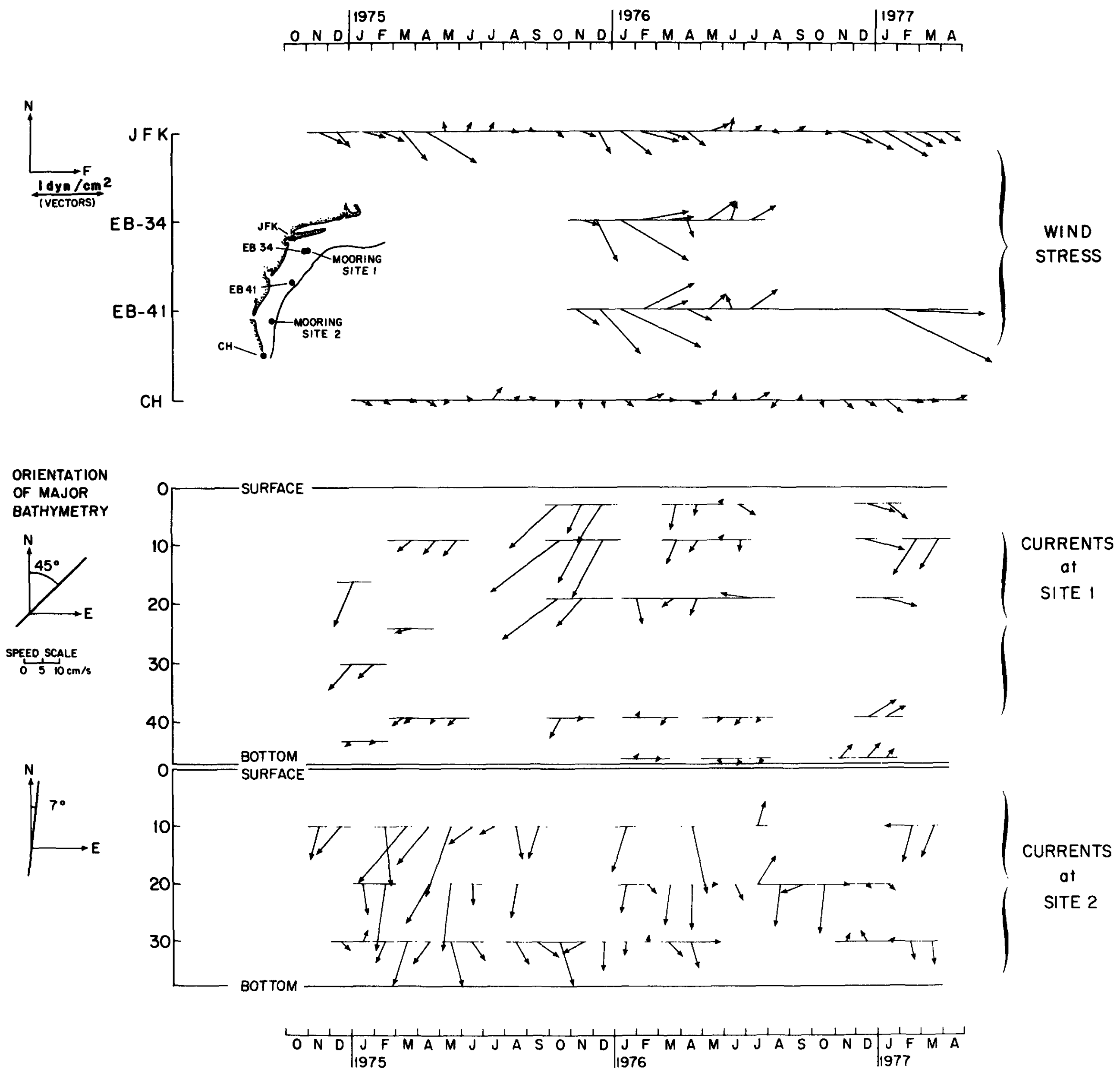

Figure 7.12 Vector time series of monthly-mean wind-stress and subsurface currents observed within the Middle Atlantic Bight during the period October 1974 to April 1977. The measurement locations are shown in the inset map. The monthly-mean wind-stress at two coastal and two mid-shelf sites are shown in the top four time series. The currents shown at site 1 are taken from Mayer, Hansen, and Ortman (1979). Boicourt and Chuang (personal communication) supplied the current data shown at site 2. The depth scale at left is shown in meters.

orientation of the regional topography at each mooring site is indicated on the left of the figure. The locations of the four meteorological stations and the two current measurement sites are also shown in figure 7.12.

This composite figure illustrates several key features of the lower-frequency current variability of the Middle Atlantic Bight. The monthly-mean wind-stress data show both a definite offshore increase in strength and a clear tendency for cyclonic veering in most months. The wind-stress vectors are generally aligned to within 20° although the wind-stress vectors at EB41 and JFK differ in orientation by 40° in February 1976. The monthly-mean currents are generally directed downshelf at all measurement levels except during a few periods when the alongshore flow is reversed for 1 to 3 months. While the magnitude of the possible errors in these current measurements is not clear [see Mayer et al. (1979)], the alongshore current components at both sites exhibit a vertical shear consistent with an offshore increase in the mean density field, so that the upshelf flow during the reversals appears to be strongest near the bottom. While strong upshelf flows occur on shorter time scales, the submonthly current fluctuations are generally comparable to or less than the long-term mean currents at both sites. Since a similar picture is presented for the nearshore monthly-mean currents off New Jersey and Long Island by EG&G (1978) and Scott (personal communication), we conclude that the monthly-mean subsurface currents over most of the Middle Atlantic Bight are directed downshelf except for relatively infrequent reversals of one to several months duration. The vertical and horizontal structure of these current reversals is clearly complex although the preliminary data shown in figure 7.12 suggest that the monthly-mean current fluctuations over the middle and outer shelf may have a large alongshelf coherence scale.

We shall now focus on the monthly-mean wind stress and currents observed during two specific periods—the spring and summer of 1976 and the winter of 1976-1977. Winds for the period May through July 1976 were more persistent and stronger than normal, and the monthly mean wind-stress vectors were directed toward the north and northeast (Diaz, 1980). This wind stress drove a definite upshelf and onshore surface flow over the nearshore region from New York to Cape Cod (Frey, 1978), which, coupled with a higher-than-average river discharge in May and June, led directly to the severe pollution of the western Long Island beaches (Swanson, Stanford, and O'Conner, 1978). This wind stress pattern also apparently caused the reduced downshelf flow observed at both mid-shelf sites (figure 7.12), and the weak upshelf near surface current in June at site 1, and the strong reversal in July at site 2. The dissolved-oxygen concentration in the near-bottom water over the shallower half of the New Jersey shelf reached very low values (less than 2 $\mathrm{ml\,l^{-1}}$) during June through August, resulting in an extensive mortality of shellfish valued at $60 million. This major anoxic event and the environmental conditions that may have caused the severely depleted oxygen levels off New Jersey and not elsewhere in the Middle Atlantic Bight are examined in the comprehensive report edited by Swanson and Sinderman (1980). We note here that Armstrong (1980), Walsh, Falkowski, and Hopkins (1980), and others attribute the 1976 anoxic event to: (a) the early development of the seasonal pycnocline [this is clearly shown in the temperature time series shown for site 1 by Mayer et al. (1979)], which reduced the initial dissolved-oxygen concentration in the deeper water and inhibited later oxygen replenishment; and (b) an excessive local oxygen (respiration) demand created by an unusual abundance of the dinoflagellate *Ceratium tripos* advected onto the New Jersey shelf by the weak upshelf and onshore deep flow. The monthly-mean current measurements for this period presented here and by Mayer, Hansen, and Minton (1980), Han, Hansen, and Cantillo (1980), EG&G (1978), Butman et al. (1979), and Ou (1979) suggest that the persistent northward wind stress caused an upshelf flow over the shallower New Jersey shelf with perhaps an offshelf and enhanced downshelf flow of shelf water over the outer New Jersey shelf, producing in essence a persistent mesoscale clockwise gyre during June. Han et al. (1980) have used a diagnostic numerical model together with observed density and current data to predict the quasi-steady current field over the New York Bight during the 1976 anoxic event. The computed deep transport fields indicate both a net convergence of deep water over the inner New Jersey shelf and strong cross-shelf flow occurring in the Hudson shelf valley.

A more dramatic case for atmospheric forcing of shelf circulation on the monthly time scale occurred during the winter period, November 1976 through January 1977. Winds during this 3-month period were both more persistent and stronger than normal, and the monthly-mean wind-stress vectors were generally directed toward the east-southeast (Wagner, 1977). This wind-stress pattern produced strong upshelf currents near the bottom at both sites (thus providing an effective bottom stress in *opposition* to the upshelf wind-stress component), and strong upshelf and offshore flow in the upper 20-30 m at site 1. The monthly-mean downshelf subsurface flow on the southern side of Georges Bank was less than average in November and December 1976, and the mean alongshelf current was essentially zero in January 1977 (Folger, Butman, Knebel, and Sylvester, 1978). The downshelf mean flow was observed over the shelf in February. These limited observations strongly imply that a sufficiently strong and persistent adverse wind-stress pattern can effectively stop and reverse the normally downshelf monthly-mean flow over much of the Middle Atlantic Bight. On December 15, 1976, the tanker *Argo Merchant* ran aground on Nantucket Shoals, and later broke apart on December 22 during a major storm, causing one of the largest oil spills off the east coast of the United States. It was indeed fortunate that almost all of the oil remained on the surface and was blown off the continental shelf by the unusually strong and persistent winds occurring during late December and early January (Grose and Mattson, 1977; Mattson, 1978).

In summary, the sparse current observations discussed here suggest that the monthly-mean currents over most of the Middle Atlantic Bight are primarily downshelf except during infrequent periods of strong and persistent adverse wind forcing. The spatial structure of the monthly-mean current field is quite complex, especially during periods of upshelf flow.

The Annual Mean Circulation We have plotted in figure 7.13 the long-term mean currents observed at four shelf sites in recent moored-array programs. Only records of duration 1 year or longer have been used, and information about the individual current measurements is listed by site number in table 7.2. The mean currents are plotted as vectors with the magnitude equal to the average speed. The rectangle shown around the head of each current vector represents the statistical uncertainty in the computed mean. Each side of the rectangle corresponds to the standard error ϵ for the current component parallel to that side, where ϵ has been computed using the formula $\epsilon = \sigma_{LF}/\sqrt{T/\tau_c}$, where σ_{LF} is the low-frequency (subtidal) standard deviation of that current component, T the length of the record in days, and τ_c the correlation time scale in days. The quantity T/τ_c is an estimate of the number of independent observations in the current time series, based on the assumption that the time series is stationary. We have used here $\tau_c = 5$ days for both the alongshore and cross-shore components. This is a conservative estimate since Mayer et al. (1979) and Flagg (1977) found somewhat shorter correlation time scales. However, the use of a more accurately determined correlation time scale would not substantially change the size of the standard errors shown in figure 7.13. Two wind-stress vectors taken from Saunders (1977) have also been plotted in figure 7.13 to illustrate the orientation of the mean wind stress with respect to the shelf break over the southern and northern sections of the Middle Atlantic Bight.

These measurements of the annual mean current field on the continental shelf clearly demonstrate that the mean subsurface alongshelf flow is directed toward the southwest through the Middle Atlantic Bight. While there are not enough long-term measurements to define the spatial structure of the mean circulation in detail, this picture when combined with other current information and hydrographic evidence substantiates the previous concept that over much of the shelf (except perhaps very near the major estuaries), the long-term subsurface currents are downshelf towards the southwest. The mean vectors at sites 1, 2, and 3 show a definite tendency for onshore veering of the mean velocity vector with increasing depth that is consistent

Table 7.2 Tabulation of Long-Term Mean Currents, Standard Errors, and Subtidal Standard Deviations Obtained at Four Sites in the Middle Atlantic Bight and Georges Bank Region[a]

Site number	Location	Orientation of topography (θ_T)	Coordinate orientation (θ_c)	Measurement time span	Record length (days)	Water depth (m)	Nominal instrument depth (m)	E′ (cm s^{-1}) Mean	E′ St. error	E′ St. dev.	N′ (cm s^{-1}) Mean	N′ St. error	N′ St. dev.
1	36°52′N, 75°03′W	7°	0°	Nov. 74–April 77	469	38	10	2.1	±1.3	13.0	−5.8	±2.5	24.7
					573		20	0.3	±0.9	9.2	−4.4	±2.1	22.3
					668		30	0.5	±0.6	6.9	−2.6	±1.5	17.5
2	39°28′N, 74°15′W	36°	36°	May 72–Dec. 76	1089	12	4.5	1.6	±0.2	2.7	−3.9	±0.8	12.2
					1082		10	−1.1	±0.2	3.0	−2.1	±0.6	8.2
3	40°07′N, 72°54′W	45°	45°	June 74–March 77	482	47	10	1.3	±0.7	7.0	−5.2	±0.9	9.0
					575		20	0.5	±0.7	7.0	−3.9	±1.1	12.0
					528		41	0.2	±0.2	2.0	−0.9	±0.5	5.0
4	40°51′N, 67°24′W	58°	58°	May 75–Dec. 78	742	85	45	−0.8	±0.3	3.9	−8.8	±0.6	7.6
					692		75	0.4	±0.3	3.4	−3.6	±0.5	5.5

a. Each velocity has been decomposed into an offshore component E′ and an orthogonal alongshore component N′ where the orientation of the alongshore direction relative to north is indicated by θ_c. The orientation of the regional topography relative to north is denoted by θ_T. Except at site 1, θ_c and θ_T are equal. [Boicourt and Chuang (personal communication), EG&G (1978), Mayer et al. (1979), and Butman and Noble (1978, 1979) provided the data for sites 1–4, respectively.]

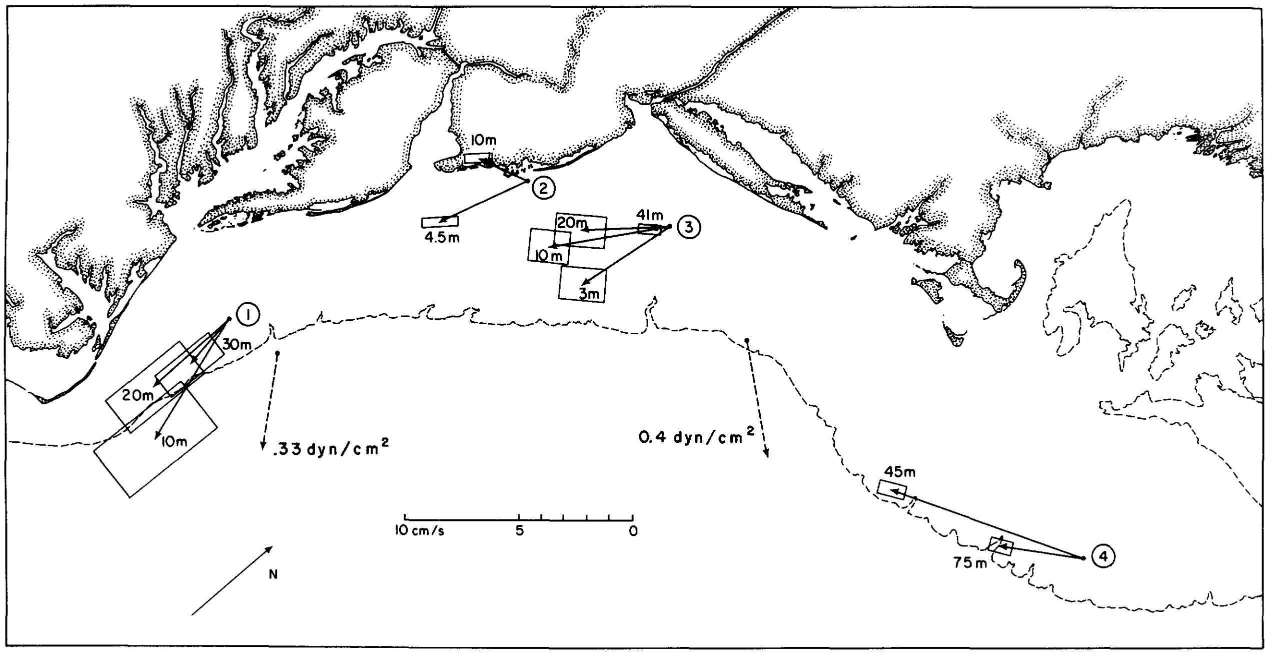

Figure 7.13 Map of long-term mean currents computed from 1-year or longer current time series with moored current meters in the Middle Atlantic Bight and Georges Bank region. The individual sites are circled and numbered according to table 7.2; the measurement depth in meters is shown next to the mean-current vector. The standard error for each mean-current computation is indicated by the rectangle around the head of the current vector. Two representative mean wind-stress vectors taken from Saunders (1977) are also shown to illustrate the relative orientation of the mean wind-stress near the shelf break.

with the mean surface- and bottom-drifter results reported by Bumpus (1973). This mean-current veering is most pronounced at site 2 off the New Jersey coast where nearshore upwelling must occur on average.

Beardsley, Boicourt, and Hansen (1976) speculated that the mean alongshore transport of shelf water through the Middle Atlantic Bight might be approximately constant throughout the year, so that the mean velocities should increase over the southern section of the Middle Atlantic Bight where the cross-sectional area of the shelf is smaller. This mechanism may account for the somewhat larger mean currents found at site 1 off Cape Henry. The observed low-frequency current variances are also largest at site 1, where the typical standard deviation of the alongshelf current component is about 20 cm s^{-1}, or a factor of two larger than the typical values found at the other sites. Although more measurements are needed to determine accurately the spatial distribution of the subtidal current variance over the Middle Atlantic Bight, the few values given here do suggest that the wind-driven-current variability is significantly larger over the shallower southern section of the Middle Atlantic Bight.

In summary, these long-term direct measurements demonstrate that the alongshelf flow, if averaged over a year or longer, is directed toward the southwest at all levels throughout the water column except perhaps near the surface, and even there the shorter-term measurements at 3 m reported by Mayer et al. (1979) suggest that the long-term mean surface flow is also directed towards the southwest with an offshore component. The long-term current measurements made on the southern flank of Georges Bank help substantiate the notion from hydrographic evidence that the southern side of Georges Bank is the immediate upstream source region for much of the shelf water found in the Middle Atlantic Bight. We shall now discuss the dynamic ideas that have been put forth to explain the observed mean circulation over the Middle Atlantic Bight.

A dynamical discussion: As mentioned above in section 7.3.3, Sverdrup, Johnson, and Fleming (1942) suggested at a very early date that the southwestward flow over the shelf was driven against friction by an alongshore pressure gradient, but this dynamic explanation was ignored by Bumpus (1973) and others as too vague and speculative, especially after the accuracy of the original coastal geodetic leveling results was questioned by Sturges (1968) and Montgomery (1969).

Some 30 years later, Stommel and Leetmaa (1972) constructed the first theoretical model for the mean winter circulation on a flat two-dimensional continental shelf driven by a steady, uniform wind stress and a distributed fresh-water source located at the coast. The model incorporated linear Ekman dynamics with constant vertical friction and diffusion coefficients, while the cross-shelf salt balance was maintained by an advective-diffusive Taylor shear-dispersion mechanism.

Stommel and Leetmaa (1972) applied their model to the Middle Atlantic Bight and concluded that an alongshore pressure gradient, or surface slope of order 10^{-7}, must exist to drive the mean alongshore flow toward the southwest against the mean wind stress, which (in their model computations) had an alongshore component toward the northeast.

Using a similar dynamic model for the steady flow over a sloping two-dimensional shelf, Csanady (1976) also concluded that an alongshore surface slope must exist to account for the observed circulation within the Middle Atlantic Bight. The essential physical argument for the alongshore pressure gradient can be found in the alongshore momentum balance. The net Coriolis force on the onshore flow is negligible since the depth-averaged onshore flow is very small, and the divergence of the onshore flux of alongshore momentum (or onshore Reynolds stress) is also negligible. Since the mean windstress and the bottom stress have alongshore components directed toward the northeast, the boundary stresses must then be balanced by an alongshore pressure-gradient force directed toward the southwest in the direction of the mean alongshore flow. Csanady (1976) noted that if the alongshore surface slope remained approximately constant across the shelf, then the total pressure-gradient force associated with this surface slope, the alongshore bottom stress required to partially balance this total pressure-gradient force, and the alongshore velocity should all increase as the local water depth increases offshore. The early mean-current data summarized by Beardsley et al. (1976) clearly show such an offshore increase in the alongshore currents. Csanady (1976) also pointed out that the alongshore pressure gradient is required to account for the line of bottom-drift divergence reported by Bumpus (1973) to occur near the 60-m isobath. Scott and Csanady (1976) then offered some supporting indirect evidence for the existence of the alongshore pressure gradient off Long Island, although their estimated value of the gradient has been questioned by Wang (1979c) and Beardsley and Winant (1979).

Csanady (1978) then presented a simple linear model for the frictionally damped steady depth-averaged flow driven over a sloping shelf by a uniform alongshore pressure gradient externally imposed at the shelf break. He found that the surface elevation over the shelf is governed by a parabolic differential equation, and that downstream from the initial (spatial) transient behavior, the alongshore surface slope or pressure gradient remained constant across the shelf. Since in this model the total alongshelf pressure gradient is balanced solely by bottom stress, the alongshore velocity increased with increasing depth. Csanady (1978) concluded that the open ocean was not "inert," but instead imposed the mean pressure gradient along the outer edge of the Middle Atlantic Bight that was required to drive the observed mean southwestward flow on the shelf.

This suggestion has been strengthened by Beardsley and Winant (1979), who found supporting evidence in the numerical simulations by Semtner and Mintz (1977) of the circulation in the western North Atlantic. Semtner and Mintz (1977) constructed a multilayered numerical model, using a rectangular basin oriented with the long side of the basin adjacent to the east coast of the United States. The model basin had a flat bottom except along the western boundary where a simple continental shelf and slope were located. The model ocean is driven by steady zonal surface temperature and wind-stress patterns, and Leetmaa and Bunker (1978) indicate that the model wind-stress curl is reasonably realistic over the northern region of the model. The model topography and wind-stress distribution are shown in figure 7.14. Semtner and Mintz (1977) first conducted an initial highly damped spin-up experiment and then two shorter experiments with reduced dissipation in which the model Gulf Stream became unstable and mesoscale eddies developed. The mean volume transport, surface pressure, and temperature fields obtained in the three experiments are quite similar in the northern shelf and slope region (the model equivalent to the Middle Atlantic Bight) and so only the mass transport and surface elevation patterns found in the spin-up experiment are shown in figure 7.14. In all three experiments, a mean alongshore pressure gradient corresponding to a surface slope of order 10^{-7} was imposed over the northern shelf and upper slope by the large wind-driven cyclonic gyre formed north of the Gulf Stream. Because continental runoff was not included in the model, Beardsley and Winant (1979) concluded that these model results demonstrated that the shelf circulation in the Middle Atlantic Bight could be considered a boundary-layer component of the offshore large-scale ocean circulation.

Beardsley and Winant (1979) also discussed (without reaching any conclusions) the possibility that a major fresh water source like the St. Lawrence system could produce a coastally trapped pressure-gradient field that might extend downstream through the Middle Atlantic Bight. This other possible driving mechanism has been recently ruled out by Csanady (1979), who concluded from the steric set-up distribution over the continental margin from Cape Hatteras to the Grand Banks that the effects of the St. Lawrence outflow are primarily confined to the Scotian shelf and the northern section of the Gulf of Maine.

The mean southwestward flow of shelf water through the Middle Atlantic Bight thus appears to be driven primarily by an alongshore pressure gradient imposed at the shelf break by the deep-water cyclonic gyre found between the continental shelf and the Gulf Stream. The magnitude of this mean alongshore pres-

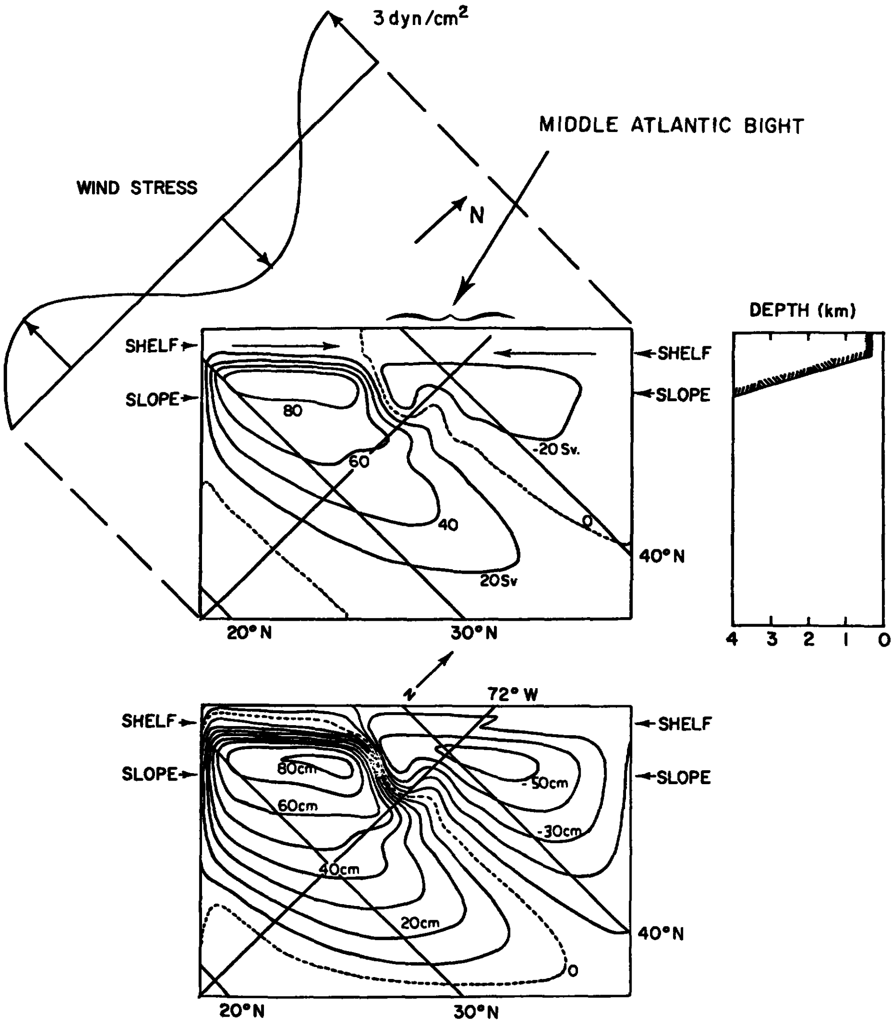

Figure 7.14 Steady mass transport (top) and surface elevation (bottom) fields obtained by Semtner and Mintz (1977) in their initial spin-up experiment. The contour intervals for top and bottom are $20 \times 10^6\ m^3 s^{-1}$ and 10 cm. The zonal wind-stress pattern and the model topography is shown to the upper left and right of top. The model equivalent of the Middle Atlantic Bight is indicated at top.

sure gradient can be expected to vary along the shelf break due to changes in both the strength of the regional wind stress over the shelf and in the relative orientation of the wind stress with respect to the regional shelf topography. The calculations of Bush and Kupferman (1980) suggest that the relative magnitude of the alongshore pressure gradient may be quite small off the southern section of the Middle Atlantic Bight where the regional mean wind stress is directed primarily offshore with a weak alongshore component toward the southwest. While local wind stress and runoff definitely influence the mean nearshore circulation, the model results of Semtner and Mintz (1977) indicate that the mean circulation over much of the Middle Atlantic Bight may be viewed as a boundary-layer component of the large-scale general circulation of the western North Atlantic. Continental runoff injects fresh water into the boundary layer, which is then mixed with an onshore flux of upper slope water to produce shelf water of intermediate salinity, while entrainment by the Gulf Stream provides the principal downstream sink of this shelf water.

What does the dynamic model presented here imply about the seasonality of currents in the Middle Atlantic Bight? Both Saunders (1977) and Leetmaa and Bunker (1978) indicate that the wind-stress and wind-stress-curl distributions vary significantly over the western North Atlantic on a seasonal basis, yet the phase of these seasonal fluctuations is not stable enough from year to year to produce a very significant annual peak in the surface wind-stress spectra at Weathership C reported by Willebrand (1978). The volume transport of shelf water through the Middle Atlantic Bight corresponds to just a few percent of the mean volume-transport estimates for the slope water gyre given by Worthington (1976) and Leetmaa and Bunker (1978). The numerical ocean-model computations made with realistic time-dependent winds by Willebrand, Philander, and Pacanowski (1980) suggest that the volume transport of the slope water gyre should *not* vary significantly over the year, and, in fact, Thompson (1977) found no evidence of an annual signal in the long-term deep-current measurements made near site D on the New England continental rise. These two results imply that the magnitude of the alongshore pressure gradient imposed at the shelf break should remain nearly constant over the year. Since the long-term current measurements made just off the New Jersey coast by EG&G (1978) also exhibit no significant annual cycle, we suggest that the very low-frequency currents over most of the Middle Atlantic Bight will reflect broadband forcing and not exhibit a significant annual variation.

7.3.4 Summary and Some Remaining Problems

We have attempted here to describe the shelf response to wind forcing on three time scales. Synoptic-scale atmospheric disturbances and in particular winter cyclones can drive strong transient current fields that tend to move along the shelf in phase with the forcing. The cross-shelf momentum balance is approximately geostrophic and both the current and subsurface pressure fluctuations are generally coherent over much of the Middle Atlantic Bight, reflecting the relatively small size of this shelf region with respect to the atmospheric forcing. The synoptic-scale shelf response appears to be consistent with continental shelf-wave theory. The direct effect of wind forcing is also evident in the monthly-mean shelf circulation, although the observed currents have a complex spatial structure and other processes like runoff and offshelf forcing contribute to the variability on this time scale. The mean flow over the Middle Atlantic Bight is primarily driven not by local runoff and wind stress but by the large-scale wind stress and heat-flux patterns over the western North Atlantic. Thus the observed currents can be decomposed into a mean component driven by a steady offshelf forcing and a fluctuating component driven by the regional wind stress field acting over the shelf.

This review has focused on the wind-driven-circulation components in the Middle Atlantic Bight. The actual influence of density stratification on the different components of the general circulation is still not clear, and only a crude estimate of the flushing rate of the shelf is available. The processes controlling the local position and movement of the shelf-slope-water front are poorly known. As noted by Fofonoff (chapter 4, this volume), satellite infrared mapping of the sea surface temperature has greatly added to our perception of the spatial variability and complexity of the near-surface current and thermal fields. The satellite infrared photograph shown here in figure 7.15 illustrates thermal structures or fronts on a wide range of scales throughout the Middle Atlantic Bight. The most pronounced thermal front is the shelf-slope-water front, and while it crudely follows the shelf break, some of the colder shelf water does extend far offshelf into the slope water or along the northern side of the Gulf Stream. Several anticyclonic Gulf Stream eddies are also shown in the slope water north of the Gulf Stream. It remains to be determined whether Gulf Stream eddies and other very low-frequency phenomena in the slope water have much real influence on the flow of shelf water through the Middle Atlantic Bight.

Appendix: Annual Air-Sea Interaction Cycles and Mean Runoff for the Middle Atlantic Bight

The mean and average monthly heat-flux cycles over the southern section of the Middle Atlantic Bight have been described by Bunker (1976). We show here his results, plus additional data kindly supplied by Bunker (personal communication) for the northern section, to give a complete picture of the seasonal heat flux cycles for the entire Middle Atlantic Bight. The monthly air and sea surface-temperature and heat-flux cycles for both sections shown in figure 7.16 and the annual mean values are listed in table 7.3. We note that the sea surface-temperature cycle lags the net heat-flux cycle by about 90°, and that the approximately 17°C increase in sea surface-temperature from March to August is consistent with a uniform mixing of the net internal-energy gain by the shelf water during that period to a mean depth of 30 m.

Average precipitation and evaporation data for the Middle Atlantic Bight are also shown in figure 7.16 and in table 7.3. The evaporation rates are computed from Bunker's heat-flux data. Although few precipitation data are available over the Middle Atlantic Bight, precipitation along the coast of the Middle Atlantic Bight is roughly uniform and exhibits relatively little seasonality [see Geraghty et al. (1973) and Lettau, Brower, and Quayle (1976)]. We thus have used data from New York City as an estimate for the precipitation cycle over the entire Middle Atlantic Bight. Mean steamflow of fresh water entering the Middle Atlantic Bight along the coast via the major estuaries is also given in table 7.4. Most of the fresh water is contributed by a few major sources, especially the Chesapeake Bay. The mean precipitation crudely balances evaporation over the Middle Atlantic Bight and the net input of fresh water via local precipitation minus evaporation into the Middle Atlantic Bight is minor in comparison to runoff at the coast. Advection of fresh water (meaning here zero salinity) from the Gulf of Maine and Georges Bank region accounts for most of the total fresh water flux into the Middle Atlantic Bight.