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PHYSICAL OCEANOGRAPHY OF CONTINENTAL SHELVES

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Introduction

Knowledge of the physical oceanography of continental shelves has increased tremendously in recent years, primarily as a result of new current and hydrographic measurements made in locations where no comparable measurements existed previously. In general, observations from geographically distinct continental shelves have shown that the nature of the flow may vary considerably from region to region. Although some characteristics, such as the response of currents to wind forcing, are common to many shelves, the relative importance of various physical processes in influencing the shelf flow field frequently is different. In the last several years, the scientific literature on shelf studies has expanded rapidly, with that for separate regions, to some extent, developing independently because of the variable role played by different physical effects. Consequently, it seems that a simultaneous review of progress in physical oceanographic research in different shelf regions would be especially useful at this time in order to help assess the overall progress in the field. This is appropriate also because the last quandrennial report did not include a review of continental shelf dynamics and much of our present understanding has been obtained in the last eight years. With the above objective, the assembler of this paper (J. S. Allen) felt that the most knowledgeable discussions would be given by those actively working in each area. The following multi-authored report has been compiled as a result. Included are sections on the physical oceanography of continental shelves, in or off of, the eastern Bering Sea, northern Gulf of Alaska, Pacific Northwest, southern California, west Florida, southeastern U.S., Middle Atlantic Bight, Georges Bank and Peru. These discussions clearly point to the diverse nature of the dominant physics in several of the regions, as well as to some of the dynamical features they share in common.

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Eastern Bering Sea
(J. D. Schumacher, T. H. Kinder, L. K. Coachman)

The largest continental shelf sea of the World Ocean, outside the Arctic, lies contiguous to the west coast of Alaska in the eastern Bering Sea (Figure 1). This shelf, which is bounded on the south by the Alaska Peninsula and on the north by Alaska and Siberia subtends 11° of latitude and exceeds 500 km in width at its narrowest point. The shelf deepens gradually from the shore to about 170 m at the shelf break which is indented by several huge canyons. The shelf connects with the Gulf of Alaska by Unimak Pass and with the Arctic Ocean through Bering Strait. Averaged over a year, a significant volume ($\sim 1 \times 10^6$ m^3/s; Coachman et al., 1975; Coachman and Aagaard, 1981) flows northward through Bering Strait, and a lesser volume ($\sim 0.15 \times 10^6$ m^3/s; Schumacher et al., 1982) of Alaskan coastal water enters through Unimak Pass. Waters above the shelf receive an excess of precipitation over evaporation [Reed and Elliott, 1979], and river discharge, principally from the Kvichak, Kuskokwim, and Yukon Rivers, adds about 1.5×10^4 m^3/s [Roden, 1967; Favorite et al., 1976]. Apparently much of the remaining transport required to make up the Bering Strait outflow comes across the shelf south of Cape Navarin. Among this shelf's unique features are its vast size, its "leak" on the northern boundary, and the seasonal production of ice.

Hydrographic Features

Prior to 1975, limited Japanese, Soviet, and American research had been carried out. Results from these studies, based primarily on broad (in both time and space) hydrographic surveys, have been summarized by Dodimead et al. [1963], Arsenev [1967], Ohtani [1973], Takenouti and Ohtani [1974], Favorite et al. [1976], and Ingraham [1981].

Since 1975 there has been a notable increase in the amount of research devoted to study of sub-regions of the shelf or specific oceanographic phenomena. This is a response to the possibility of petroleum development, the substantial harvests of pollock and crab, and the need to predict ice

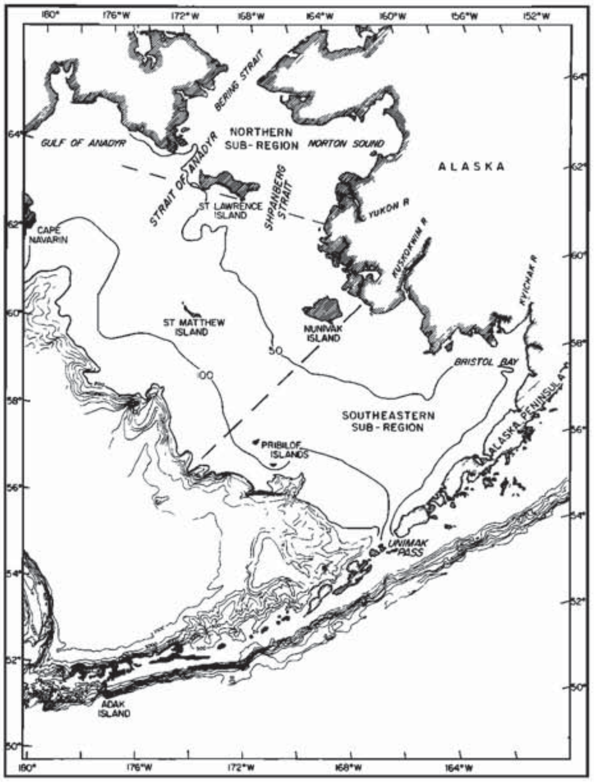

Fig. 1. Bathymetry of the Bering Sea basin and shelf (contours are in meters). Geographic names used in the text are also shown [after Sayles et al., 1979].

movement. The research has focused on the southeastern shelf and on that portion of the shelf around and north of St. Lawrence Island, including Norton Sound [Kinder, 1981]. Kinder and Schumacher [1981a] synthesized recent hydrographic work over the southeastern shelf. The shelf is divided into distinct domains, delineated by water depths (z) and separated by fronts (Figure 2). Within the coastal domain, $z < 50$ m, tidal mixing exceeds buoyancy input, and the water (away from the direct influence of river discharge) is mixed vertically. In the middle shelf domain, $50 \text{ m} \leq z \leq 100$ m, when the seasonal input of buoyancy (either from melting ice or insolation) exceeds tidal mixing, two-layered structure obtains. Separating these domains is the inner front, the zone of transition in the balance between tidal mixing and buoyant energy input [Schumacher et al., 1979]. The change in water column structure crossing the 50-m isobath was noted previously by Ohtani [1973] and by Muench [1976]. Even during winter, when surface cooling and increased frequency and strength of storms destroys the two-layered nature of the middle shelf waters, there is a stronger density gradient across the 50-m isobath than over the middle shelf [Schumacher and Kinder, 1983].

Within the middle shelf there is little or no significant advection [Kinder and Schumacher, 1981b; Schumacher and Kinder, 1983] so that heat content is dictated by air/sea exchange [Reed, 1978], and the salt flux required to maintain the nearly constant mean salinity appears to be tidally driven diffusion [Kinder and Coachman, 1978; Coachman et al., 1980]. These recent studies support the suggestion by Takenouti and Ohtani [1974] that the cold (<0 to 3C) bottom layer of the middle domain observed in summer is primarily formed _in situ_. At the seaward edge of this domain (~100 m) the bottom slope becomes five times greater than is typical for the majority of the shelf. Here the water column undergoes a change over a broad (about 50 km wide) transition zone, the middle front.

The outer domain, $100 \text{ m} \leq z \leq 170$ m, is characterized by well mixed upper and lower layers separated by an intermediate layer containing much finestructure. The outer shelf is bathed by slope waters which are warmer and more saline than the waters of the middle shelf. This juxtaposition of water masses of slightly differ-

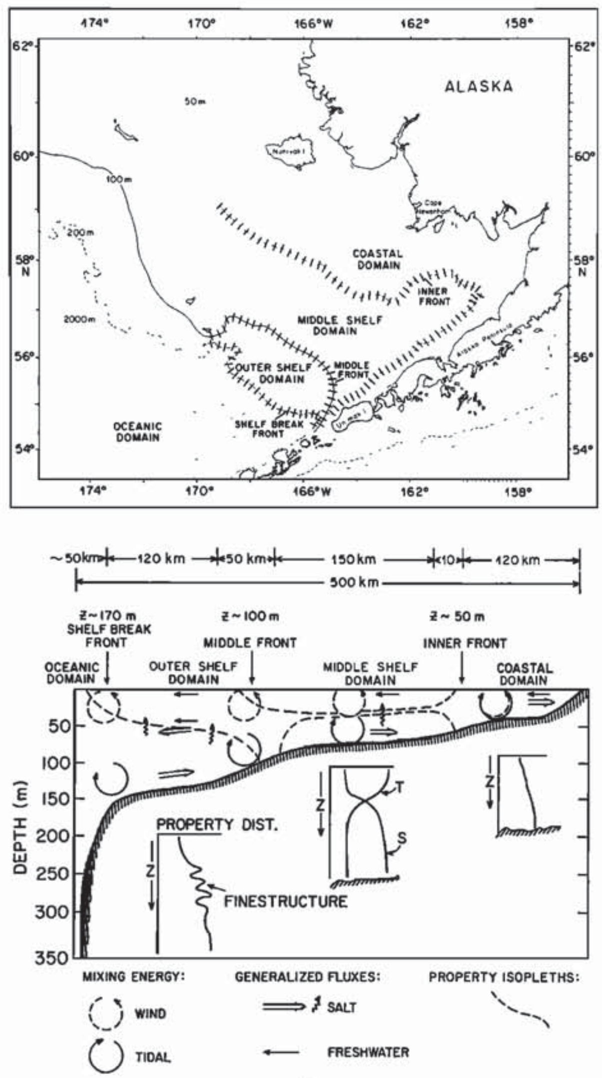

Fig. 2. Approximate locations of domains and fronts over the southeast Bering Sea shelf [after Kinder and Schumacher, 1981a] (top) and a schematic interpretation in the cross-shelf plane of energy balance, fresh and salt water fluxes and vertical structure [after Coachman, et al., 1980] (bottom).

ent densities results in lateral interactions with interleaving of water masses occurring at vertical scales of 1 to 25 m within the mid-water column [Coachman and Charnell, 1977, 1979]. Outer shelf waters intrude shoreward near the bottom while middle shelf waters extrude seaward above them, and within the middle front vertical fluxes appear to be enhanced [Coachman et al., 1980; Coachman and Walsh, 1981].

Kinder and Coachman [1978] described the shelf break front, which separates the outer shelf from the oceanic domain, and recognized its essentially haline character; it is the zone of relatively steep upper-layer salinity gradients (about 0.5 g/kg in 50 km) dividing upper layer shelf water from the upper-layer water of the Bering Slope Current. Paralleling the shelf break from near Unimak Pass to near Cape Navarin, this current transports $\sim 5 \times 10^6$ m^3/s northwestward [Kinder et al., 1975]. The oceanic region immediately seaward of the shelf break front is replete with mesoscale eddies [Kinder and Coachman, 1977; Kinder et al., 1980].

The other sub-region of the eastern Bering Sea shelf which has been recently studied in detail is the northern shelf (north of about 62°N), including Norton Sound. Coachman et al. [1975] report three identifiable water masses. The most saline water mass (Anadyr) lies west of St. Lawrence Island and on the west side of Bering Strait. The least saline waters (Alaskan Coastal) lie contiguous to the Alaskan Coast on the east, and between these a separately definable water mass of intermediate salinity (Bering Shelf) is present. More recent studies [Aagaard et al., 1981; Schumacher et al., 1983] support the observations of the zonal salinity gradients differentiating these water masses. They also have shown that during ice formation brine rejection results in salinity values as high as 35 g/kg in local areas. Observations from Norton Sound indicate that in summer the water column is strongly two-layered in both temperature and salinity with the eastern portion of the Sound apparently being isolated from the western part [Muench et al., 1981].

The large region of the eastern Bering Sea shelf that lies between the southeastern and northern sub-regions has been less well studied. However, hydrographic data collected during summer [Kitani and Kawasaki, 1978] and winter [Bourke and Paquette, 1981; Salo et al., 1980; Muench, 1983] suggest that the three domains present over the southeastern shelf extend northward to the vicinity of St. Lawrence Island, and the three separate water masses of the northern region are identifiable with these domains.

Currents

Over the southeastern shelf, tides dominate the kinetic energy of the water, often comprising 90% of the fluctuating kinetic energy. Farther north, however, the tides become less energetic and are a much smaller percentage of the total fluctuating kinetic energy. Pearson et al. [1981] examined current and water pressure records from the eastern Bering Sea shelf and produced cotidal charts of the M_2, N_2, K_1, and O_1 constituents. They found that the M_2 constituents (the largest) vary from 35 cm/s along the Alaska

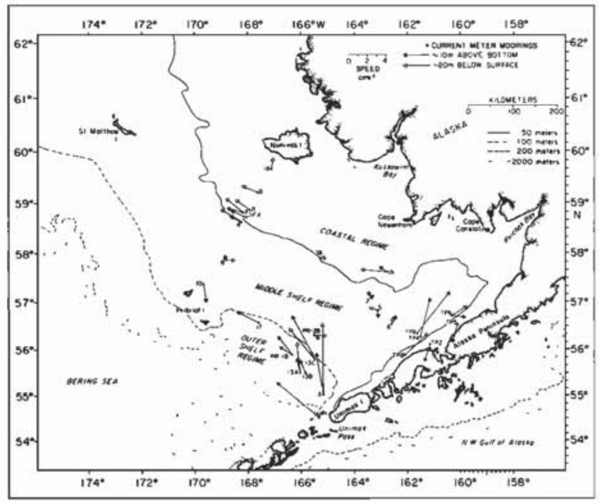

Fig. 3. Mean flow based on all records at each mooring site [from Schumacher and Kinder, 1983].

Peninsula to 3 cm/s or less in Norton Sound, where there is an amphidrome. The tidal wave enters the Bering Sea from the North Pacific Ocean through the central and eastern passages of the Aleutian Islands and then propagates eastward onto the shelf. The semidiurnal tide propagates as a Kelvin wave along the Alaska Peninsula, but appears to be converted to a Sverdrup wave upon reflection in inner Bristol Bay. The diurnal tides cooscillate between the deep basin and the shelf with amphidromes located between Nunivak Island and the Pribilof Islands and west of Norton Sound. Recent numerical models by Hastings [1976] and Sunderman [1977], which are vertically integrated models of the M_2 component over the entire Bering Sea shelf, and by Leendertse and Liu [1977] and Liu and Leendertse [1978, 1979], which are three-dimensional models of all significant constituents, have all predicted tides. Pearson et al. [1981] found generally good agreement between observations and the models of Sunderman and of Leendertse and Liu.

More than twenty record-years of direct current measurements have been collected over the southeastern shelf since 1975, with somewhat fewer observations from the northern shelf and Norton Sound. Three mean and low-frequency current regimes have been identified over the southeastern shelf (Figure 3), and these regimes are nearly coincident with the previously described hydrographic domains [Kinder and Schumacher, 1981a,b; Schumacher and Kinder, 1983]. Coastal waters from the Gulf of Alaska shelf flow into the Bering Sea through Unimak Pass and then apparently continue northeastward along the Alaska Peninsula [Schumacher et al., 1982]. Within Bristol Bay the flow becomes counterclockwise and then follows the 50 m isobath past Nunivak Island and continues northward. Currents appear to be strongest near the inner front, paralleling this feature at speeds between 1 and 6 cm/s, with the highest speeds occurring in winter. Although fluctuating kinetic energy is dominated by tides (about 96%), there are significant pulses of flow which are wind driven. However, a combination of baroclinic geostrophic

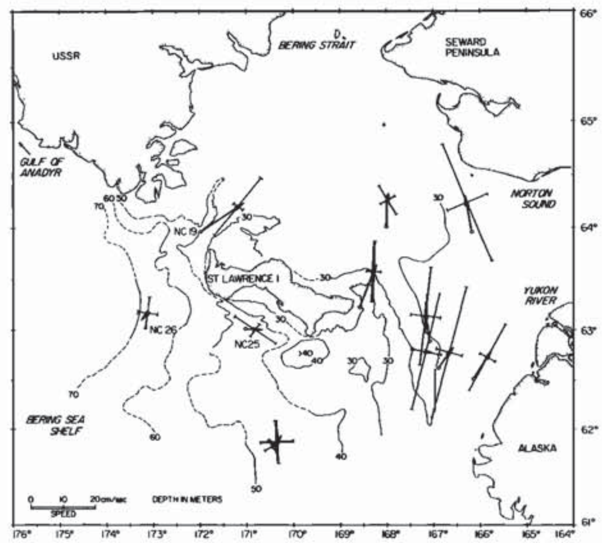

Fig. 4. Vector mean current with 1-standard deviation bars along major and minor axes shown on the end of the current vectors [from Schumacher et al., 1983].

flow and residual current produced by the interaction of the tides with shoaling bathymetry appears to account for the observed mean velocities [Schumacher and Kinder, 1983].

The middle shelf current regime, which is bounded by the inner and middle fronts, is a region where there are wind-driven pulses, but the mean current is statistically insignificant (<1 cm/s) except near the regime boundaries. Kinetic energy at periods which correspond to meteorological forcing (2 to 10 days) is of similar magnitude to that in the coastal domain. Due to the great width of this shelf, however, there are no coastal boundaries within a Rossby deformation radius, and Ekman divergence at the coast does not appear to be an important mechanism for generating currents. Instead, currents respond to the wind as rotating vectors. It is this lack of organized flow, together with the seasonal development of an upper layer which acts as an insulating lid, which permits the bottom layer of the middle domain to retain its cold signature throughout the summer.

The outer shelf regime lies between the middle and shelf break fronts. Vector mean flow is statistically significant, with along-isobathic speeds between 1 and 10 cm/s (toward the northwest) and across-isobathic speeds between 1 and 5 cm/s (northeastward). Unlike the two shoreward regimes, the higher speeds are not necessarily a winter feature. Because the cross-shelf flow does not typically extend onto the middle shelf, the middle front is often a region of convergence, and in the outer domain advection is equal to or more important than tidal diffusion in achieving cross-shelf fluxes [Coachman, 1982]. Estimates of baroclinic geostrophic velocities and those generated by topographic rectification of tides on the larger bottom slope under the middle front are similar in magnitude to the observed along-isobathic flow [Schumacher and Kinder, 1983]. The outer regime is richer than the two other regimes in kinetic energy at periods greater than 10 days, which may be a result of propagation of energy from the Bering Slope Current and its eddies landward across the outer shelf.

The Bering Slope Current is a mixture of Alaskan Stream/Bering Sea water [Takenouti and Ohtani, 1974; Favorite et al., 1976] with the former source entering the Bering Sea through passes in the western Aleutian Islands [Reed, 1971; Favorite, 1974; Swift and Aagaard, 1976; Sayles et al., 1979]. Measured and inferred speeds [Kinder et al., 1980] and model currents [Han and Galt, 1979] suggest an along-slope flow that averages 5 to 15 cm/s toward the northwest.

Circulation over the northern shelf is dominated by a generally northward flow of water bound for the Arctic Ocean. This pattern can be temporarily reversed because of large-scale meteorological forcing, particularly in early winter [Coachman and Aagaard, 1981]. Recent current measurements have shown that mean and low-frequency flow is usually aligned with isobaths (Figure 4); both east and west of St. Lawrence Island and through Bering Strait mean flow often reaches 10 to 15 cm/s or more [Coachman et al., 1975; Schumacher et al., 1983]. South and southwest of the island, the flow is much weaker. In Norton Sound, the northward mean flow appears only in the western portion [Muench et al., 1981], and mean currents in the remainder of the sound are weak, although wind-driven currents with instantaneous speeds up to 100 cm/s have been observed [Drake et al., 1980; Cacchione and Drake, 1982]. The few direct measurements made between the southeastern and northern sub-regions of this shelf appear to confirm delineation of the flow field into three regimes, as on the southeastern shelf.

Climatology

A major influence on the general atmospheric circulation over the Bering Sea is the region of low pressure normally located in the vicinity of the Aleutian Islands, referred to as the Aleutian Low. This feature is a manifestation of the passage of storms, which dominate climatology of the Bering Sea [Brower et al., 1977; Overland, 1981]. During winter there is a tendency for two storm tracks, one parallel to the Aleutian Islands and one curving northward along the Siberian coast; however, there is always a decrease in the number of storms with increasing latitude [Overland and Pease, 1982]. Due to the juxtaposition of the Aleutian Low and the Siberian High, the mean winter winds over this shelf are from the northeast, and outbreaks of cold polar air, which continue for one to two weeks, are a common winter phenomenon. The winter mean winds are stronger than those during summer and have been shown to have an interannual signal which conditions both sea surface temperature [McClain and Favorite, 1976] and ice coverage [Niebauer, 1980, 1981, 1983; Overland and Pease, 1982]. During summer, storms tend to migrate northward into the Bering Sea, and mean winds are from the south.

Ice

While the strong winter winds have been shown to result in stronger sub-tidal flows over much of the southeastern shelf [Kinder and Schumacher, 1981b; Schumacher and Kinder, 1983], they also have a dramatic impact on water temperature and ice production. Ice cover is a seasonal feature of the eastern Bering Sea shelf, varying from none in summer to greater than 80% coverage of 0.5 to 2.0 m thick ice during its maximum extent in March [Niebauer, 1980; Pease, 1980]. Fay [1974], Muench and Ahlnäs [1976], Pease [1980], McNutt [1980, 1981], and Aagaard et al. [1981] have discussed the importance of the northern Bering Sea as an area where ice is preferentially produced along south-facing coasts because of wind-driven surface divergence. The ice produced in polynyas (areas of open water) is exported southward under the prevailing northerly winds, eventually melting at the southern marginal ice zone. Salinization during freezing has a measureable effect on both flow and water properties in the vicinity of polynyas [Schumacher et al., 1983]. At the southern ice boundary, bands of ice which accelerate away from the pack have been identified in satellite photographs [Muench and Charnell, 1977; McNutt, 1980, 1981]. The melting ice causes a vertical stratification in the water column, which may produce a baroclinic geostrophic flow along the marginal ice zone [Niebauer, 1982; Muench, 1983]. Niebauer [1980, 1981] and Overland and Pease [1982] showed that the interannual variations in ice extent can be hundreds of kilometers, and these variations are generally correlated with either the wind field or the location of storm tracks.

Physical-Biological Interactions

Studies of biological processes over the southeastern Bering Sea and in the vicinity of the ice edge have followed the progress in understanding of the physical processes. Because of the inherent interrelationship between the biology and the physics, biological studies have also contributed to an understanding of physical processes. Hattori and Goering [1981] studied nutrient distributions and primary production and found a zonation that was congruent with the three hydrographic domains. Iverson et al. [1979, 1980] and Goering and Iverson [1981] demonstrated that the phytoplankton and zooplankton communities of the ecosystem were organized by the system of fronts and domains. In quantitative assessments, they inferred relative rates of diffusion and advection of the biological variables based on physical hypotheses about the vertical and horizontal rates of diffusion. Coachman and Walsh [1981] constructed a diffusion model for the cross-shelf nutrient flux using a combination of physical, chemical, and biological data. Kinder et al. [1983] showed that the location of tidal fronts around the Pribilof Islands are correlated with the distribution of feeding seabirds. Niebauer et al. [1981] and Alexander and Niebauer [1981] showed that the atmospheric forcing of ice edge location can influence phytoplankton distributions, especially through effects on water column stability.

Northern Gulf of Alaska
(T. C. Royer)

The weather over the continental shelf in the northern Gulf of Alaska provides large seasonal signals in temperature, wind, pressure and precipitation. The consequence is that the meteorological forcing over this shelf is not subtle. Shelf response here is further complicated by the existence of two coastal driving forces, wind stress and runoff, instead of the more typical case where wind stress acts alone.

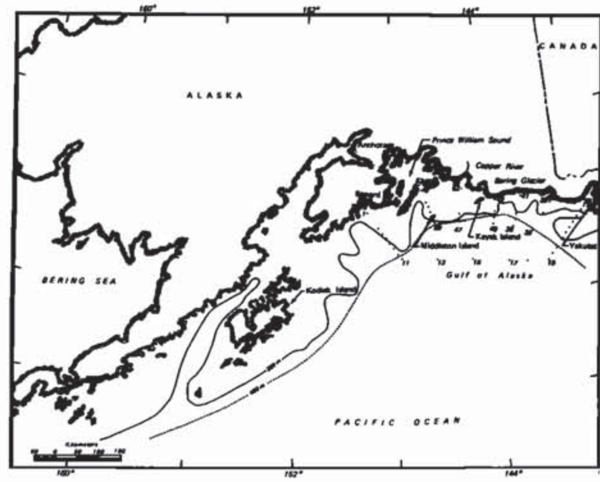

Fig. 5. Northern Gulf of Alaska where early (1974-78) Outer Continental Shelf studies occurred.

The continental shelf in the northern Gulf of Alaska is quite broad, up to 200 km, and contains numerous deep troughs (Figure 5). The northern shelf has two major islands (Kayak and Middleton) and farther to the west Kodiak Island dominates the mid-shelf area. West of Kodiak Island, the Aleutian Islands form the boundary between the North Pacific Ocean and Bering Sea. The shelf commonly has depths in excess of 200 m beginning within several kilometers of the mountainous coastline. Twenty percent of the coastal region is covered with glaciers and the only major river is the Copper River.

Initial circulation studies in the Gulf of Alaska were carried out in the 1920's and 1930's [McEwen et al., 1930; Thompson and Van Cleve, 1936; Thompson et al., 1936], but there followed a long period (1936-1970) when few observations were made. Geostrophic computations indicated that there was a general westward flow over the shelf in the northern Gulf of Alaska with occasional eastward velocities.

The bulk of the work in the Gulf of Alaska between 1930 and 1970 has dealt with the hydrological description of the offshore waters. As part of this pre-1970 work, several review papers of the offshore waters were published [Dodimead et al., 1963; Favorite et al., 1976; Filatova, 1973; Ingraham et al., 1976]. Additional descriptions of the flow in the Alaska Current/Alaska Stream system were done by Favorite [1967] and Roden [1969], while the system was treated theoretically by Thomson [1972] and Veronis [1973]. The Alaska Current, which flows westward along the shelf break, is included here in the context of a possible offshore influence for the coastal circulation as speculated by Favorite [1974] and Reid and Mantyla [1976].

Meteorological Forcing

The measurement of meteorological parameters in the Gulf of Alaska has been limited by lack of suitable observing stations. Most studies requiring wind data use those determined by the Fleet Numerical Oceanographic Central and/or upwelling indices determined by Bakun [1973] using the same pressure data set. When compared with actual winds on Middleton Island [Livingstone and Royer, 1980], Bakun's winds were reasonably accurate in displaying the seasonal trends but missed the high frequency changes. From studies of the coastal meteorology near Yakutat, it was found that orographic effects are extremely important for nearshore meteorology, with winds increasing significantly nearshore [Reynolds et al., 1978].

The winds over the Gulf of Alaska are dominated by strong cyclonic systems in the winter, leading to westward alongshore winds, and by a weak high in summer. These lows are enhanced by cyclogenesis [Winston, 1955] and tend to be trapped by the coastal mountains. Adiabatic ascent of the moist marine air over the coastal mountains leads to extremely high precipitation rates in the coastal drainage areas [Royer, 1982]. Hydrologists have generally neglected these drainage areas because of the lack of large river networks, but average discharges for some of the major rivers have been presented by Roden [1967].

Circulation

The circulation of the deep water (>100 m) over this shelf responds to the seasonal wind stress with the renewal of bottom water in the local fjords taking place in late summer [Muench and Heggie, 1978]. The absence of the strong wind field in summer allows a relaxation of downwelling and the onshore intrusion of relatively warm, salty water from the central Gulf of Alaska [Royer, 1975]. This type of deep water renewal is supported by the seabed drifter studies near Kodiak Island where some drifters released near the shelf break were found within the local bays [Ingraham and Hastings, 1974; Muench and Schumacher, 1980].

Prior to 1974, only one moored current observation had been reported for the Gulf of Alaska [Pearson, 1973]. That mooring deployed near Middleton Island indicated weak westward flow along bathymetry with reversals. Recent current measurements in the northeastern Gulf of Alaska indicate that the flow generally follows the local isobaths in the nearshore regions [Hayes and Schumacher, 1976]. However, away from the coast near Icy Bay (west of Yakutat Bay) the eddy kinetic energy increases offshore as one approaches the shelf break and Alaska Current [Hayes, 1979]. Toward the west, the mean energy of the current increases, but its relative eddy kinetic energy decreases [Niebauer et al., 1981]. Therefore with weaker flow in the northeast Gulf of Alaska, eddies play a more important role than in the northwestern part. While the eddies located along the shelf break seem to be transient, a permanent eddy has been located to the west of Kayak Island [Galt, 1976]. This island

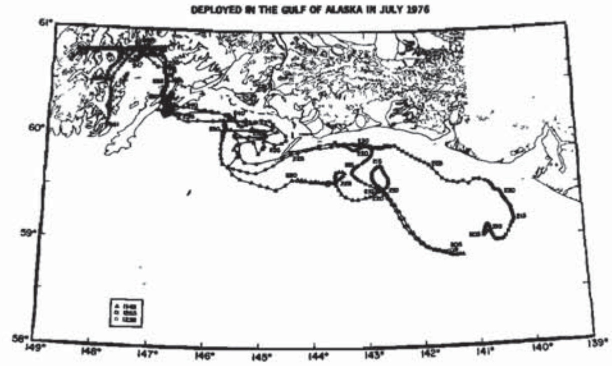

Fig. 6. Trajectories of satellite-tracked drifting buoys illustrating the general alongshore flow and the eddy west of Kayak Island.

plays an important role in the coastal and offshore flow as it deflects a portion of the relatively fresh coastal current southward into the Alaska Current. West of Kayak Island, the Alaska Current can be clearly identified by a low salinity surface layer [Ingraham, 1979], by anomalous sea surface temperatures [Royer and Muench, 1977] and occasionally by suspended particulate matter concentrations [Feely et al., 1979]. In the vicinity of Kodiak Island and Cook Inlet, Muench et al. [1981] have described the circulation of lower Cook Inlet as a continuous channel flow connected with the Gulf of Alaska coastal region north of Kodiak Island. Mysak et al. [1981] have analyzed fluctuations in the current of Shelikof Strait and Lagerloef [1983] has investigated the topographic control of the flow off the southeastern tip of Kodiak Island.

Other current measurements have been made with drifters released in the central Gulf of Alaska and recovered along the Alaska Peninsula [Favorite and Fisk, 1971]. More recently, satellite-tracked drifting buoys that were released south of 48N in the vicinity of 160W, entered the Alaska Current, traversed an eddy near Sitka and then entered the coastal current. They progressed northwestward along the coast [Kirwan et al., 1978]. The drifter that traveled farthest ran aground near Icy Bay.

Additional satellite-tracked drifting buoys were released near Yakutat and Icy Bay in 1976 to describe possible trajectories of pollutants from oil and gas development. In general, these buoys wandered over the outer and mid-shelf regions, slowly progressing shoreward until they became involved with the coastal current (Figure 6). At that point, they progressed westward following the coastline. They traveled around Kayak Island and entered the permanent eddy west of the Island. After leaving the eddy, the drifters entered Prince William Sound where they became stranded [Royer et al., 1979]. These trajectories describe the general coastal circulation in this area, except that the flow continues out of Prince William Sound to the west. Three drifters released in the Alaska Current in deep water near Kodiak Island in 1978 did not move shoreward, but instead made a circuit around the Alaskan gyre before one of them traversed the shelf and went ashore west of Kodiak Island [Reed, 1980].

Overview

The wind plays a major role in determining the nearshore circulation. The low pressure system over the central Gulf of Alaska during the winter creates a coastal convergence and downwelling with westward alongshore flow [Royer, 1975]. However, the maximum coastal current transport ($\geq 1 \times 10^6 m^3/s$) precedes the maximum wind stress by several months. This may be caused by local runoff [Royer, 1981b]. Both of these effects can be seen in the annual signal in coastal sea level [Royer, 1979; Reed and Schumacher, 1981]. Apparently, the coastal flow is isolated from the effects of the Alaska Current which either has no seasonal signal [Reed et al., 1980] or has a maximum transport in late spring or early summer [Royer, 1981a]. The relatively fresh water in the Alaska Coastal Current should mix offshore across the shelf as it moves alongshore, its rate of cross-shelf movement being dependent on friction [Kao, 1981]. However, even hundreds of kilometers from its sources the coastal flow remains quite narrow, less than 25 km wide [Schumacher and Reed, 1980; Royer, 1982]. This implies that something is opposing this spreading tendency. It is speculated here that the wind acts in a manner to maintain the coastal current as a narrow flow adjacent to the coast. Thus, the wind and freshwater effects may combine, probably in a nonlinear fashion, to maintain the coastal flow.

Throughout the central Gulf of Alaska, the nutrient concentrations below the halocline are among the highest in the world's ocean [Reid, 1961]. Downwelling and runoff keep these nutrients at depth and offshore so that a relaxation of downwelling or a decrease in the runoff will allow an upward displacement of this nutrient-rich water. Thus, the absence of winds or precipitation might enhance coastal primary production, in contrast with upwelling regions where active wind forcing is necessary for high production.

The temporal variability of the wind field in the Northeast Gulf of Alaska can create situations where the coastal current locally disappears altogether [Muench and Hachmeister, 1982]. Schumacher et al. [1982] indicate that the coastal current is an important source of low salinity water for the Bering Sea. Other implications of this coastal current have yet to be investigated such as the interaction of the Alaska Coastal and Alaska Currents, the effects of the coastal current on the circulation of adjacent fjords and estuaries, and the exchange of water between the North Pacific and Bering Sea by the flow through the Aleutian passes. While extensive sampling of the physical parameters has taken place at a few selected sites, the alongshore variability of the coastal current has not been investigated. The use of chemical and biological parameters as tracers for this current seem promising for future studies.

Pacific Northwest
(A. Huyer, R. L. Smith)

The Pacific Northwest coast runs nearly north-south from Cape Blanco at 43N to Cape Flattery at 48N (Figure 7). There are only a few large estuaries and bays, and the continental shelf topography is quite uniform compared with other U.S. shelf regions. The shelf width varies from about 15 to 40 km with the shelf break occurring at a depth of about 200 m. Currents over the shelf have been measured, primarily with subsurface-moored Aanderaa current meters, for varying periods in different years. The longest records were obtained near 44°45'N. Many of the results of observations made before 1978 have been summarized in review papers by Huyer and Smith [1978], Allen [1980], and Huyer [1983].

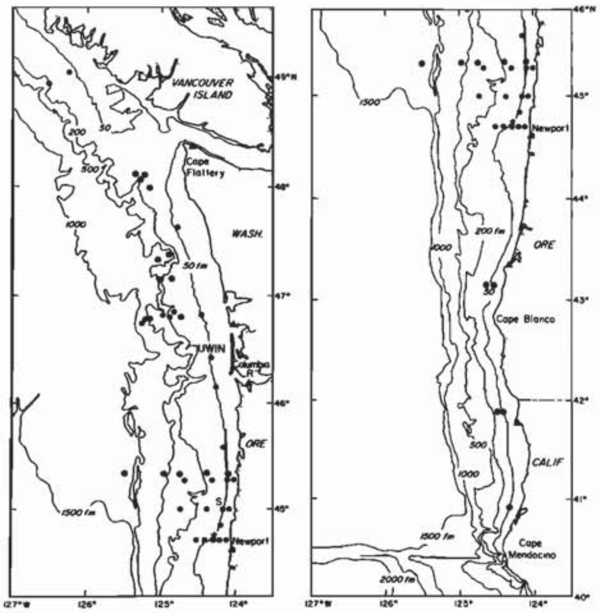

Fig. 7. Locations of current meter moorings (dots) maintained for a month or more and tide gages (triangles) along the Pacific Northwest coast. The star at 44°45'N indicates the position of Poinsettia, a mooring maintained continuously for more than a year.

At low frequencies (f < 0.6 cpd) (i.e. periods longer than 40 hours), the currents are primarily alongshore, with the alongshore component typically exceeding the onshore component by a factor of two or more. The fluctuations have very high vertical coherence [Smith, R. L., 1981], and the alongshore fluctuations are in phase at different depths. A quasi-barotropic first empirical orthogonal function accounts for more than 80% of the fluctuation energy in each season. These low frequency fluctuations are coherent and nearly in phase with local wind and sea level. They are also coherent and in phase across the width of the continental shelf [Huyer et al., 1978], and are coherent over large alongshore separations; Hickey [1981] showed that simultaneous low-passed current measurements separated alongshore by 480 km had linear correlation coefficients greater than 0.7. This high correlation was observed in spite of the fact that the alongshore separation crossed some major canyons, including Astoria Canyon and Juan de Fuca Canyon.

Typical amplitudes of the alongshore current fluctuations vary from 10 cm/s in summer to

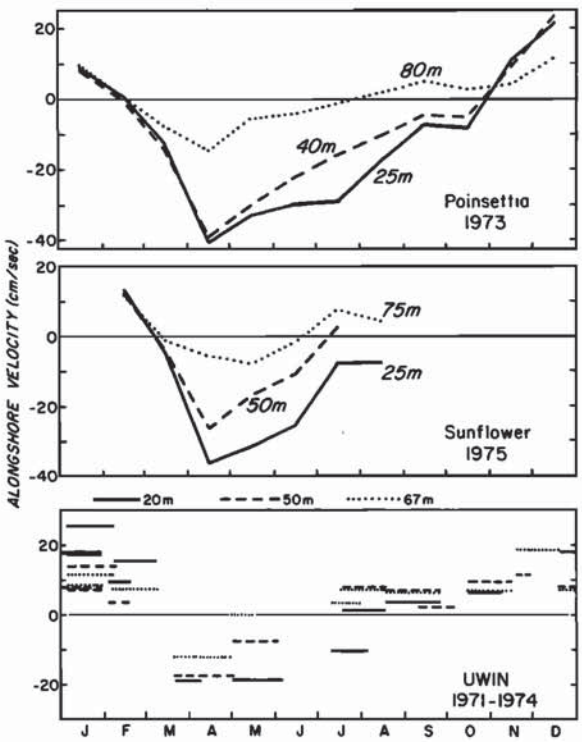

Fig. 8. Monthly mean values of the alongshore component of the currents at Poinsettia in 1973 and at Sunflower in 1975. Also shown are the record lengths and the value of the mean alongshore current for each record at UWIN, from Smith, Hickey and Beck [1976]. The locations of Poinsettia, Sunflower(S) and UWIN are shown in Figure 7. Currents were rotated by -20° at Poinsettia, -13° at Sunflower, and +13° at UWIN.

30 cm/s in winter. These dominate over the tidal and inertial currents which both have amplitudes of ∿5 cm/s [Torgrimson and Hickey, 1979; Kundu, 1976]. However, the converse is true for the onshore component: low frequency fluctuations with typical amplitudes of <3 cm/s [Huyer et al., 1978; Smith, R. L., 1981] are obscured by the roughly isotropic tidal and inertial currents. Thus, observations of limited duration often suggest complex onshore/offshore circulation patterns [e.g., Johnson and Johnson, 1979], while averages from longer period observations usually suggest a simpler circulation pattern (e.g. Smith, R. L., 1981). The low frequency fluctuations in the onshore/offshore component are only partly due to variations in the alongshore wind stress [Allen and Smith, 1981]. The remainder might be due to changes in the alongshore pressure gradient. These fluctuations result in a net onshore eddy heat flux in summer [Bryden et al., 1980].

Probably the most distinctive feature of currents over the shelf of the Pacific Northwest is the strength and the repeatability of the seasonal cycle of the alongshore currents. In each year when measurements over mid-shelf are available, mean monthly currents for December, January and February are northward at all depths; mean monthly currents for April and May are southward at all depths; and mean monthly currents in June and July are southward at the surface (Figure 8). The amplitude of the seasonal cycle of the surface currents is comparable to the amplitude of the strongest low frequency fluctuations [Huyer et al., 1979]. Thus, the direction of the surface current is variable in winter when the fluctuations are strongest, but nearly constant in summer when fluctuations are weaker. The mean current is northward at all depths in winter, southward at all depths in spring, and southward at the surface but northward near the bottom in summer [Huyer et al., 1978].

The low frequency current fluctuations are much weaker in summer than in winter or spring, as would be expected from the weaker fluctuations in wind stress [Huyer et al., 1978]. In summer, the bottom stress is weak, and can be neglected in the vertically integrated momentum equations [Allen and Smith, 1981]. In winter and spring, however, the bottom stress is much larger, and Hickey and Hamilton [1980] have shown that the bottom stress is an important term in the momentum equations in those seasons.

The vertical gradient of the alongshore velocity, i.e., the shear, also varies seasonally. In winter, there is almost no mean vertical shear. In both spring and summer, the vertical shear is negative, i.e. currents are more strongly southward near the surface than near the bottom. In winter the shear fluctuates, but in spring and summer the shear is almost constant [Huyer et al., 1979]. Associated with this change in the vertical shear is a change in the density distribution. In winter, isopycnals are nearly level and have the greatest variability near 100 m over the mid-shelf. In spring, the mean isopycnals slope upward toward the coast and the greatest variability is at the surface near shore. In both the velocity field and the density field, the difference between the winter and spring mean values is greater than the standard deviation in either season [Huyer et al., 1979].

The onset of the typical spring-summer regime, with southward surface currents, sloped isopycnals and persistent shear, occurs very quickly, over a day or two, during a single upwelling event [Huyer et al., 1979]. Sea level falls quickly during the same event. This rapid spring transition seems to be due partly to the local wind stress [Huyer et al., 1979], partly to the large scale winds along the coast, and partly to the alongshore pressure gradient [Werner and Hickey, 1982] which is probably southward at this time of year [Hickey and Pola, 1982].

The fall transition (i.e. the disappearance of the southward surface current, the decay of the persistent shear, and the return to level isopycnals) seems to occur very gradually. A poleward undercurrent appears over the shelf during the summer [Hickey, 1979; Halpern et al., 1978; Huyer et al., 1979]. This undercurrent increases in strength while the southward surface current weakens during August and September. The shear seems to decrease in almost a step-wise fashion, with a step at each northward wind event. The water column is gradually restratified, as isopycnals lose their upward slope. This process appears to occur first at the surface, and only later at depth. Again, changes in the stratification appear to occur in association with northward wind events.

One question that continues to intrigue investigators is the nature of the relationships among the seasonal cycles in the alongshore current, the wind and sea level. One school feels that the seasonal cycle in the currents is forced primarily by the local wind, while other schools argue that it is forced primarily by winds at distant locations, or by the alongshore pressure gradient. On seasonal time scales, alongshore currents and coastal sea level are strongly correlated. The spring zero crossing in sea level (referred to its long-term mean) occurs prior to the zero crossing in wind or wind stress. The converse is observed in autumn. These lags might imply that local wind stress is not sufficient to explain the seasonal cycle in sea level. However, if wind and wind stress are also referred to their long-term means before the comparison with sea level, the three variables have nearly simultaneous mean-crossings in both spring and fall. This would imply that purely local wind stress forcing cannot be ruled out. Thus, the controversy over the dynamics of the forcing of these strong annual cycles is by no means resolved.

Southern California
(C. D. Winant)

The continental shelf between the Mexican border and Los Angeles is very narrow, ranging in width between 3 and 10 km, and very steep, with bottom slopes of order 10^{-2}. The Gulf of Santa Catalina, which separates the shelf from the deeper Pacific Ocean, is characterized by depths of order 1 km with some deeper basins as well as a number of islands. The coastline is markedly curved, running practically east to west just south of Los Angeles, and north to south in the vicinity of San Diego. Current and temperature observations have been made in this area more or less continuously since 1977. Intensive observations of the cross-shelf structure of shelf circulation were made on a transect perpendicular to the coast off the city of Del Mar, California [Winant and Bratkovich, 1981] and the longshore structure of the flow was explored with an array of current meters deployed along the mid-shelf isobath (30 m) over a distance of 60 km, north from Del Mar [Winant, 1983]. Those observations are summarized in the following.

Thermal Structure

The temperature field over the southern California shelf undergoes a major seasonal change. Because density is mostly controlled by temperature here, these changes correspond to seasonal fluctuations in the density stratification which in turn affect the dynamics.

Vertical profiles of mean, maximum and minimum temperatures in different water depths along the Del Mar transect are presented for each season in Figure 9. During the summer, the thermal structure can be idealized as consisting of three layers: a shallow warm layer rarely extending more than 5 m below the surface, a colder isothermal layer near the bottom and, between these two layers, a 20 m thick thermocline. Typical temperature differences across

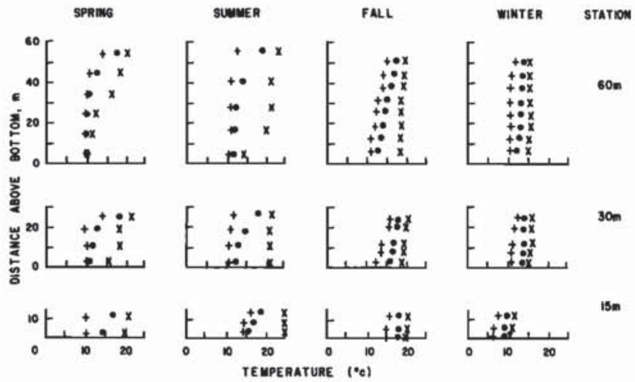

Fig. 9. Mean (•), maximum (x) and minimum (+) temperatures at each depth and during each deployment [from Winant and Bratkovich, 1981].

the water column in 60 m depth are 10°C. In the winter this mean temperature difference is reduced to the order of 2°C, and the temperature field on the shelf is approximately uniform with mean values typically between 13°C and 14°C.

During the summer, the spatially averaged temperature over the shelf is close to 15°C. Thus, the important seasonal change in the temperature field is mostly a redistribution of heat through the water column more than a net change, with colder near bottom temperatures and warmer near surface temperatures occurring during the summer.

Currents

The current field is naturally decomposed into longshore and cross-shelf components. These were observed along the Del Mar transect during six-week periods in each season except winter, when the observations lasted for three months. Spectral characteristics [Winant and Bratkovich, 1981] suggest a decomposition of the current field into mean currents, low frequency fluctuations including frequencies up to 0.6 cpd, intermediate frequency fluctuations with periods ranging between 0.6 cpd and 6 cpd and including motions corresponding to tides, inertial currents as well as currents driven by diurnal winds, and finally high frequency (>0.6 cpd) currents.

Mean Currents. Vertical profiles of longshore currents averaged over the period of the deployments are presented in Figure 10, for each season and in different total water depths. When the water column is well stratified, as it is in spring and summer, the mean currents at the surface and near the bottom are opposed. In other seasons the vertical structure is quasi-barotropic, sheared in the vertical, but without reversals. Near the surface, mean currents are always directed to the south with maximum amplitude in excess of 10 cm/s, in the winter. Cross-shelf currents averaged over such long periods are always within error of zero.

Low Frequency Currents. Low frequency cross-shelf currents are within error of zero, but longshore currents are very energetic with variance levels ranging between 60 cm^2/s^2 in the summer and 110 cm^2/s^2 in the winter. An empirical orthogonal function analysis of the variance in this band reveals that over 90% of the kinetic energy is described by two eigenfunctions in all seasons.

The largest eigenfunction typically represents 80% to 90% of the total variance in this band, and its vertical structure is quasi-barotropic. The second largest function, which is baroclinic, represents typically 5% to 10% of the energy and reaches maximum amplitude relative to the total variance in the spring and summer, when the stratification is largest [Winant and Bratkovich, 1981].

A low frequency synthetic bottom pressure was made by summing sea level observations and atmospheric pressure. There was a clear correlation between the largest current eigenfunction and synthetic bottom pressure, suggesting that in this band the cross-shelf momentum balance is geostrophic. The correlation between wind stress and currents is not significantly different from zero, although large wind events induce an important response in shelf waters [Winant, 1980]. The lack of correlation between currents and winds should mostly be ascribed to the relative absence of strong atmospheric forcing over the Southern California Bight, in comparison with other coastal areas along the U.S. seaboard.

During the summer, the alongshore coherent length scale of currents has been estimated from simultaneous current observations made along the 30 m isobath at various spacings ranging between 2.5 and 57.5 km [Winant, 1983] to be of order 30 km. This is relatively short compared to the scales in excess of 200 km observed on the Washington-Oregon shelf [Hickey, 1981] or off Peru [Smith, R. L., 1978].

Intermediate Frequency Currents. Currents in this frequency band (0.6 to 6 cpd) account for nearly half the energy found in the low frequency band, with variance levels ranging between 32 cm^2/s^2 in the winter and 60 cm^2/s^2 in the spring. The cross-shelf component of current is comparable in magnitude to the longshore component during the stratified seasons. In contrast to observations made on the eastern seaboard of the United States, currents in this band are not well correlated with the sea surface elevation, which is to say that the tidal currents have a more baroclinic than barotropic nature.

The vertical structure of the most energetic longshore eigenvector in this frequency band is quasi-barotropic [Winant and Bratkovich, 1981]. Although there is a concentration of energy near the surface during spring and summer, the amplitude is practically depth independent in the other seasons. The largest eigenvector of the cross-shelf current component is always baroclinic with a reversal which occurs near the middle of the water column.

The cross-shelf currents are coherent with the most energetic longshore mode at the mid and outer shelf locations, with a 90° phase shift between longshore currents and near surface cross-shelf currents. Cross-shelf currents in turn are coherent with the temperature, an index of isotherm displacement, with a phase lag of 270° in the upper half of the water column. These phase relations are consistent with a baroclinic mode standing in the cross-shelf plane.

During the summer, the longshore coherent length scales associated with currents in this band are mostly even shorter than the length scales associated with low frequency currents, on the order of 5 to 10 km. Near bottom longshore currents show evidence of a wave-like fluctuation propagating poleward with phase speeds of 2 m/s. The short length scales associated with the other current components and their significant kinetic energy suggest that motions in this band make an important contribution to dispersion on the shelf.

High Frequency Currents. High frequency internal waves are commonly observed during spring and summer over the southern California shelf, filling the spectral gap between tidal currents and surface gravity waves. In the upper half of the water column there is a 180° phase difference between cross-shelf currents and temperature. In addition these fluctuations are found to progress coherently from the edge of the shelf inshore at phase speeds characteristic of internal waves, but the longshore extent of these is found to be less than 2.5 km.

West Florida
(W. Sturges)

The continental shelf off the west coast of Florida is long and broad. It is about two-thirds as long as the California coast, and is as long as the distance from the Strait of Juan de Fuca to Cape Mendocino, or from Cape Hatteras to Boston. This review will focus primarily on work which has appeared since 1978. Some relevant earlier work is the compilation by the SUSIO group (1975). Also, the report by New England Coastal Engineers [1982] includes a valuable summary of older observations using drift cards.

Initial studies involved comparisons between wind forcing and the response of sea level at coastal tide gauges [Cragg et al., 1983; Marmorino, 1982, 1983a]. Because the shelf is so wide, the observed response at the coast is larger by approximately a factor of four than off the coast of Oregon, for the same wind strength. The majority of the near shore response appears to be directly forced by wind. Little energy has yet been observed associated with free shelf waves.

Observed Currents

The "shelf dynamics program" of 1973-75 provided the earliest set of current meter observations. These are predominantly on the outer shelf [Koblinsky and Niiler, 1980; Niiler, 1976]. Most of the recent observations are on the northern half of the shelf. Mitchum and Sturges [1982] found that the wind-forced currents in depths of approximately 45 m or less are highly coherent with local wind. The dominant balance is between wind stress and bottom stress. A model using a linear bottom-drag law [e.g., Scott and Csanady, 1976] fits the data well with a bottom friction parameter ("r") of 0.01-0.02 cm/s. However, because the tidal flow is small, the bottom stress is not quite linear in low frequency velocity. Mitchum and Sturges [1983] have shown that the response of sea level at the coast is to a power of wind stress less than one at St. Petersburg. They concluded that the cause was

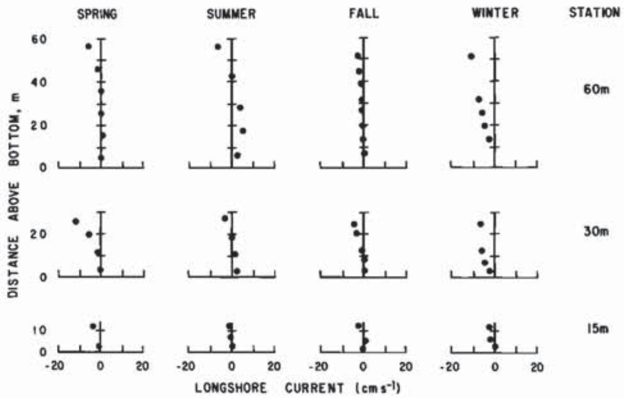

Fig. 10. Mean longshore currents. Positive values are poleward [from Winant and Bratkovich, 1981].

the result of a bottom stress law that departs from linearity in the case of small tidal flows. Marmorino [1983a,b] has reported on results from recent current meter observations, finding that the response to a 0.5 dy/cm^2 alongshore wind stress--as measured at the coast--is approximately 20 cm/s off Cedar Key but approximately 40 cm/s to the north where the shelf narrows. One finding by Marmorino seems particularly interesting. At a mooring offshore from a sharp bend in the coastline, the "longshore" currents followed the coastline direction remarkably well, in the downstream direction. Direct pressure measurements show that the pressure field decays offshore with an e-folding scale of ∿160 km, and allows calculations of geostrophic currents which are in good agreement with observed flow.

Another long record is the one obtained by Hopkins and Schroeder [1981]. They report on a pair of moorings at ∿30 m depth on the "Florida Middle Grounds" (∿25N), from Oct 78 through Nov 79, with most of the data ∿4 m above the bottom. One 4-month installation also had a current meter 10 m below the surface. They found typical speeds ∿20 cm/s in the near-bottom records associated with the passage of weather systems. The instruments also observed the effects of Hurricane Frederic which passed ∿225 km west of the mooring.

Loop Current Forcing

Observations from satellites have shown that the Loop Current comes near the edge of the shelf break; some part of the surface expression of the Loop Current comes up on the shelf [Huh et al., 1981]. The satellite observations are clearly going to be our most prolific source of data [Maul et al., 1978; Vukovitch et al., 1979], although Sturges and Evans [1983] found that the position of the Loop Current as determined from hydrographic data may be 100 to 200 km farther to the south than when determined from satellite data.

Niiler [1976] postulated a series of eddies along the shelf break--on the basis of current-meter mooring data. It seems likely that the Loop Current may force such a response at the shelf break. However, a model of Hsueh et al. [1982] is able to produce eddies at the shelf break which are similar to those suggested by Niiler; the model used by Hsueh et al. [1982] contains no Loop Current forcing, but is forced by wind. Cooper's model (see following section) suggests that many eddies on the shelf are strongly baroclinic. It is obvious that some large eddies may break off from the Loop Current and be observed partly on the shelf. Nevertheless, all such eddies on the shelf or shelf break need not be caused by the Loop Current.

The question of forcing of longshore flow on the shelf by currents in deep water--here, the Loop Current--has been addressed by Sturges and Evans [1983]. They found that sea level at coastal stations, even inside a very wide shelf, is coherent with the northernmost position of the Loop Current at periods from 8 months to 30 months. They conclude that the Loop Current is forcing longshore flows on the shelf, but they have no supporting direct observations.

Numerical Models

The model of Hsueh et al. [1982] is vertically integrated, linear, and barotropic. It can hindcast with reasonable accuracy the fluctuations of sea level observed at the coast, and it also makes shelf break eddies, as discussed above. Marmarino [1982] describes a similar, steady model. A somewhat different one (also linear) has been reported by Cooper et al. [New England Coastal Engineers, 1982]. They include stratification, vertical variation of horizontal currents and eddy viscosity. Their model reproduces coastal elevations fairly well, but it is difficult to say whether these results differ significantly from those of Hsueh et al. [1982] at describing the observed currents, as the current data available for comparison were brief. Weatherly and Van Leer [1977] showed that stratification is important to include if the effects of the bottom boundary layer are to be modelled accurately. It is difficult for this reviewer to tell whether the limitations in the models are necessarily caused by inadequately prescribed physics or whether they are due largely to our inability to describe the forcing to sufficient accuracy.

Tides

Several authors [Koblinsky, 1979, 1981; Mitchum and Sturges, 1982; Marmorino, 1983; Battisti and Clarke, 1982a] have studied tides on the West Florida Shelf. In the southern, narrower section of shelf from Naples to Tampa the tides are not strong, but north of Tampa to Apalachicola, where the shelf is much wider, semidiurnal tides and tidal currents increase significantly as a result of shelf tidal resonance [Battisti and Clarke, 1982b]. A substantial part of the energy in the observed records is from the tides, even along the narrower section of the shelf where they are weaker.

General Comments

It seems appropriate to point out that the Ocean Thermal Energy Program (OTEC) has provided

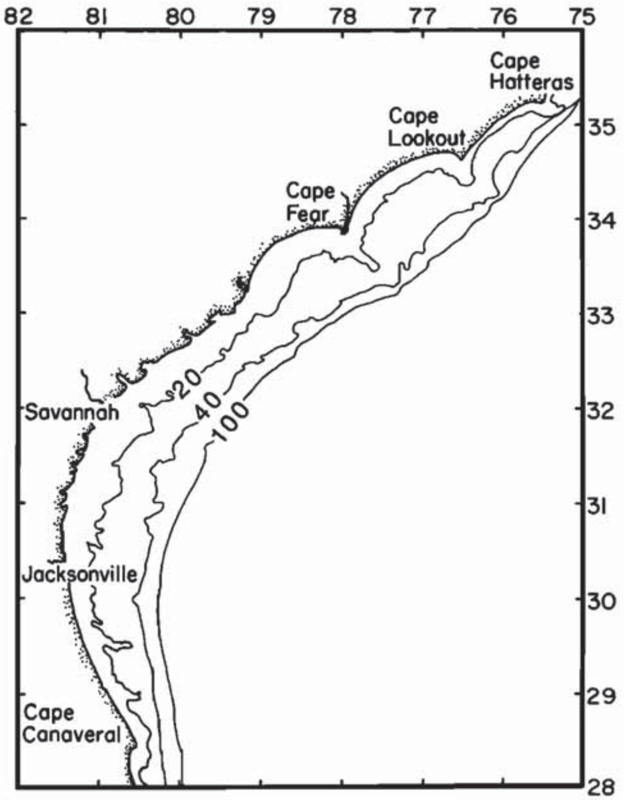

Fig. 11. Shoreline configuration and bottom topography for the continental shelf of the southeastern U.S. Isobaths are in meters.

a set of long term data [Molinari and Mayer, 1982]. Although the mooring was in deep water just off the shelf, it is relevant to studies of the shelf. Mooers and Fernandez-Partagas [1976] have given a very useful description of atmospheric frontal systems that are the major forcing mechanisms on this shelf. A new observational program on the southern part of the shelf funded by the Minerals Management Service is getting underway. The plans call for five moorings, between 26N and 28N, to be in place for approximately one year beginning January 1983.

Southeastern U.S.
(J. O. Blanton)

The continental shelf of the southeastern U.S. which stretches from Cape Hatteras to Cape Canaveral (Figure 11) is called the South Atlantic Bight (SAB). The maximum shelf width is 120 km off Savannah, Georgia, and tapers to a minimum width of 30 km off Cape Hatteras and 50 km off Cape Canaveral. The depth at the shelf break varies from 75 m in northern areas to about 50 m in southern regions. Several small rivers discharge along the shoreline over a distance of several hundred kilometers from South Carolina to northern Florida. Throughout this review, we define inner shelf as locations with depth (D) $0 < D < 20$ m; middle shelf, $20 < D < 40$ m; and outer shelf $D > 50$ m.

Factors Affecting Circulation

Saunders [1977] and Weber and Blanton [1980] have calculated wind fields for the SAB. The monthly mean wind fields throughout a given year usually have three distinct regimes: 1) winter, during which winds blow off the continent; 2) summer, when winds blow toward the north alongshore; and 3) autumn (mariner's fall), when winds blow southward alongshore. The drifter data in the SAB are consistent with the monthly mean wind patterns [Bumpus, 1973; Weber and Blanton, 1980]. The more energetic variations about the monthly mean wind stress occur primarily at frequencies of 0.1 to 0.5 cpd (the "weather band"). Mid-shelf currents and coastal sea level correlate significantly with the alongshore wind stress fluctuations [Lee and Brooks, 1979; Lee and Atkinson, 1983].

River runoff forms a band of low salinity water adjacent to much of the coast in the SAB [Atkinson et al., 1978; Atkinson et al., 1983]. This band produces an inner shelf frontal zone along 400 km of the SAB which influences the vertical distribution of inner shelf currents [Blanton, 1981; Blanton and Atkinson, 1983].

Perturbations in the western edge of the Gulf Stream influence currents on the outer shelf. As much of the total variability in Gulf Stream transport occurs within a particular season as occurs from season to season [Niiler and Richardson, 1973]. Considerable effort has been devoted in the last five years to studying the characteristics of Gulf Stream perturbations, their time scales, and the relationship of the currents on the outer shelf to these disturbances.

The interaction of Gulf Stream meanders with bottom topography influences the circulation regime on the outer shelf. In mid-shelf regions, the alongshore wind appears to be the principal mechanism affecting currents whereas, on the shallow inner shelf, wind-generated currents and tidal currents are dominant. The review will also cover those theoretical and experimental studies of the Gulf Stream that address the effects of the Gulf Stream impinging on the continental shelf.

Outer Shelf Circulation

Wave-like perturbations on the western edge of the Gulf Stream are carried northward by the Gulf Stream. Satellite data [Legeckis, 1975, 1979] suggest that waves occur each 5-10 days and propagate northward between Cape Canaveral and Cape Hatteras at about 30 km/ day. The amplitude of the waves increases abruptly from less than 15 km southward of 32N latitude to 40 km northward of 32N [Bane and Brooks, 1979]. The increase results from semi-permanent deflections of the Gulf Stream by a ridge in the bottom topography offshore of the shelf break [Bane and Brooks, 1979; Pietrafesa et al., 1978; see also Chao and Janowitz, 1979; Chao et al., 1979]. Certain manifestations of the waves have been described as shingles [Von Arx et al., 1955], meanders [Webster, 1961a] and spin-off eddies [Lee, 1975; Lee and Mayer, 1977]. Recent studies have examined these waves in considerable detail [Vukovich et al., 1979; Lee et al., 1981; Chew, 1981; Lee and Atkinson, 1983].

Recent studies have confirmed Webster's [1961a] initial description of the skewed asymmetry of the meanders [Brooks and Bane, 1981; Bane et al., 1981] and have also confirmed earlier conclusions by Webster [1961b] and Oort [1964] that energy from the fluctuating meanders is transferred to the mean Gulf Stream. Moreover, an analysis of Webster's original data [Blanton, 1975] suggested that meanders contained enough available potential energy to drive upwelled water across the shelf break as well as provide energy for feedback into the mean flow. Under certain hydrographic conditions in summer, this upwelled water can intrude along the bottom and displace large amounts of shelf water [Blanton, 1971; Blanton and Pietrafesa, 1978; Atkinson and Pietrafesa, 1980; Smith, N. P., 1981; Hofmann et al., 1981; Atkinson et al., 1980]. It passes shoreward beneath a southward-flowing warm filament on the surface [Lee et al., 1981; Bane et al., 1981].

Cold cyclonic eddies often form on the western edge of the Gulf Stream when it meanders offshore. These eddies entrain a warm streamer of surface Gulf Stream water [Lee et al., 1981; Bane et al., 1981; Lee and Atkinson, 1983]. These streamers are only 20-30 km wide, but can extend along the shelf break 10-200 km and form the shingle structure originally observed by Von Arx et al. [1955]. The streamers elongate rapidly and dissipate over the outer shelf [Lee et al., 1981]. They form an effective mechanism for mixing shelf and Gulf Stream waters.

Evidence exists that the streamers (or filaments) are larger north of latitude 32N and can develop an anticyclonic circulation [Pietrafesa and Janowitz, 1980; Chew, 1981]. The larger scale presumably results from larger meander amplitudes north of this latitude [Bane and Brooks, 1979]. Detailed measurements of the flow field within those frontal disturbances do not exist and interpretation of Eulerian measurements can be difficult [Pietrafesa and Janowitz, 1979a].

Variability of currents and temperature along the outer shelf at 0.1-0.5 cpd is primarily produced by these propagating meanders or frontal eddies. Lee and Atkinson [1983] and Lee et al. [1981] have described in detail the response of surface and bottom shelf currents to passage of frontal eddies. These disturbances are observed to transport northward momentum and heat back to the Gulf Stream, in agreement with the studies cited above on transfer of eddy energy to the mean flow. Upwelling in the cold core of frontal eddies provides the major source of nutrients for primary production on the outer shelf [Lee et al., 1981; Lee and Atkinson, 1983].

Topography around Cape Canaveral and the Carolina capes plays an important role in intensifying intrusions of Gulf Stream water along the continental shelf [Blanton et al., 1981; Leming, 1979]. The temperature of bottom waters near the Capes is anomalously low in regions where isobaths diverge [Janowitz and Pietrafesa, 1983]. A bottom temperature front separates the intruded oceanic water from ambient shelf water. The intrusion results from waves that propagate poleward along the cyclonic edge of a boundary current. Oceanic water is pulled upward through the wave trough and spreads shoreward along the shelf bottom into regions where flow on the shelf diverges. Bottom temperature data that support this scheme are shown in Figure 12.

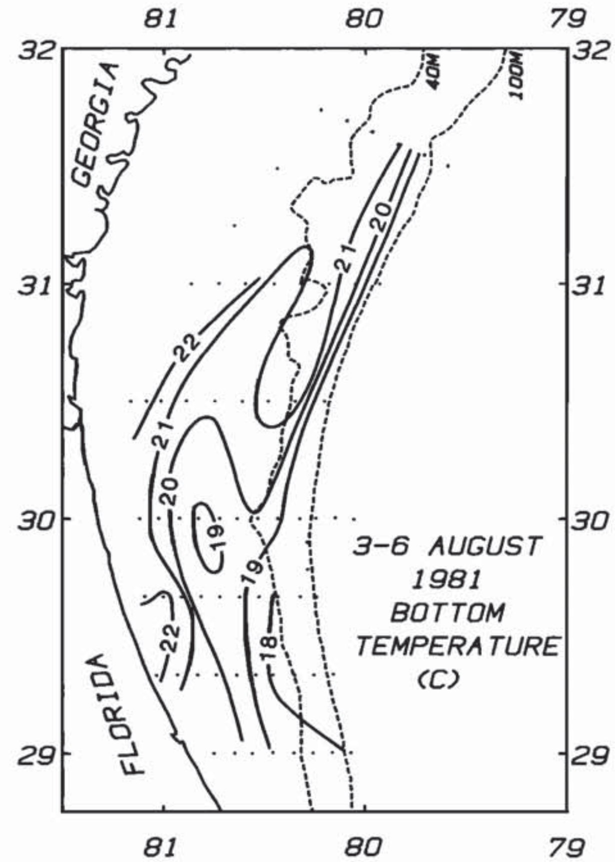

Fig. 12. Bottom temperature distribution showing deformation of the front due to passage of a wave of the cyclonic edge of the Gulf Stream. Cape Canaveral is located at 28.5N [figure courtesy of L. P. Atkinson].

Middle Shelf Circulation

Inshore from the shelf break, the predominant forcing shifts from Gulf Stream to wind. Atkinson et al. [1983] caution that the mean of cumulative weekly averages of current meter observations only stabilize after 3 or more months. Therefore, monthly means should be interpreted with caution. Mean currents along isobaths at mid-shelf calculated over the wind regimes of summer (July-August) and autumn (September-October) essentially follow the winds, i.e., northward in summer, southward in autumn [Atkinson et al., 1983].

Analysis of long term mean flows at mid-shelf suggests the alongshelf momentum balance is between an alongshelf pressure gradient forcing the flow northward and an opposing bottom stress [Lee and Brooks, 1979; Atkinson et al., 1983]. This pressure gradient may result from the northward slope downward of sea level in the Gulf Stream [Sturges, 1974] and plays a role similar to (although in the opposite direction from) that found for the open ocean along the Middle Atlantic Bight [Scott and Csanady, 1976; Csanady, 1978; Beardsley and Winant, 1979].

Low frequency variability in mid-shelf

depends mainly on local wind forcing at periods between 2 days and 2 weeks [Lee and Brooks, 1979; Lee et al., 1981; Lee and Atkinson, 1983]. Alongshore currents are coherent over at least 400 km separation with small phase lag. The mid-shelf currents in this band are essentially barotropic [Lee et al., 1982]. The alongshore currents are geostrophically balanced by cross-shelf pressure gradients and are driven by local alongshore winds. The shallowness and great width of the shelf appear to isolate mid and inner shelf regions from propagating waves along the shelf break [Lee and Brooks, 1979; Pietrafesa and Janowitz, 1980] except where the shelf is narrow off Florida [Brooks and Mooers, 1977].

Inner Shelf Circulation

The entire SAB inner shelf is affected at sometime during the year by an influx of low salinity water derived from river discharges. A band of low salinity resides essentially throughout the year off Georgia and South Carolina [Atkinson et al., 1978].

Runoff from rivers north of Cape Hatteras is occasionally advected southward around Cape Hatteras during peak discharge and episodes of southward wind stress in spring [Stefansson et al., 1971; Singer et al., 1980]. In autumn, southward wind stress advects low salinity water found off Georgia toward Florida [Atkinson et al., 1983].

Baroclinic coastal currents are set up in the frontal zone that result when low salinity water overrides the ambient shelf water. Tidal energy is insufficient to vertically mix the water column. During seasons when winds blow southward, the frontal zone is relatively narrow and steep, and the thermal wind equation suitably predicts vertical shear through the frontal zone [Blanton, 1981]. During spring off Georgia, northward wind stress seems to spread the zone farther offshore and low salinity water is lost from the inner shelf [Blanton and Atkinson, 1983]. The loss of low salinity water in spring can be explained as a response to upwelling favorable winds which decrease the horizontal density gradient and cause an additional loss of freshwater due to upper layer Ekman flux [Pietrafesa and Janowitz, 1979b].

Coastal currents measured over long time periods on the inner shelf are only now being studied [Schwing et al., 1983]. The many inlets off South Carolina and Georgia result in complex orientation of the principal axes of tidal currents [Blanton, 1980]. Vertical shear in tidal currents has been studied by Kundu, Blanton and Janopaul [1981].

The multi-inlet shoreline in the central SAB provides routes of exchange between estuaries and sound and the inner continental shelf [Kjerfve et al., 1978; Wang and Elliott, 1978]. A significant amount of sea level fluctuations in estuaries is due to sub-tidal oscillation in the open ocean off the SAB.

Inner shelf currents in summer off South Carolina in 10 m of water are predominantly longshore northward in agreement with summer wind stress [Schwing et al., 1983]. Longshore currents and winds are highly coherent at 2-12 days and currents lag winds by about 3 hours. It is puzzling that coastal sea level significantly lags longshore currents by 15 hours at frequencies of about 0.2 cpd.

Middle Atlantic Bight
(W. C. Boicourt, R. C. Beardsley)

The Middle Atlantic Bight is the curved section of the eastern U.S. continental shelf stretching between Cape Hatteras at the south and Cape Cod and Nantucket Shoals at the northeast (Figure 13). The New York Bight is the subsection of the Mid-Atlantic Bight located between the New Jersey and Long Island coasts. The shelf topography within the Mid-Atlantic Bight is relatively simple and smooth, with a generally monotonic increase in depth from the 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. This shelf region is about 800 km long and typically about 100 km wide between shore and shelf break except near Cape Hatteras where the width is about 50 km, and in the New York Bight where the shelf is about 150 km wide. While a few submarine canyons penetrate the outer shelf, the Hudson Shelf Valley off New York is the only pronounced drowned river channel to cross the shelf proper.

The first general description of the seasonal mean circulation in the Mid-Atlantic Bight is provided by Bumpus [1973] in his summary of an extensive 10-year program of drift-bottle and sea-bed drifter releases and of historical data. He concluded that a mean southwestward alongshelf flow of order 5 cm/s occurs between Cape Cod and Cape Hatteras, except nearshore during periods of strong northward winds and low runoff, and that the Nantucket Shoals and Cape Hatteras appear to be oceanographic barriers that in some sense limit the alongshelf flow. Near Cape Hatteras, the alongshelf flow turns seaward and becomes entrained in the Gulf Stream.

Since 1970, extensive current and hydrographic measurements have been made in the Mid-Atlantic Bight. These new direct current observations have led to new kinematic and dynamical descriptions of the current field and its variability. Preliminary results from some of these studies have been presented in reviews by Beardsley et al. [1976], Beardsley and Boicourt [1981], and Csanady [1982]. Partially because of the limited spatial and temporal sampling characterizing the early moored array experiments, these reviews focused on simplified quasi-two-dimensional descriptions of the seasonal mean circulation and low frequency current variability. More recent work shows that the flow field can exhibit more spatial structure and alongshelf variability than was evident in the earlier moored array observations.

We present here a brief overview of the regional hydrography and circulation within the Mid-Atlantic Bight, including recent work on spatial and temporal flow variability. The reader is referred to Williams and Godshall [1977] and Ingham [1982] for further summaries and bibliographies of historical and recent physical studies in the Mid-Atlantic Bight.

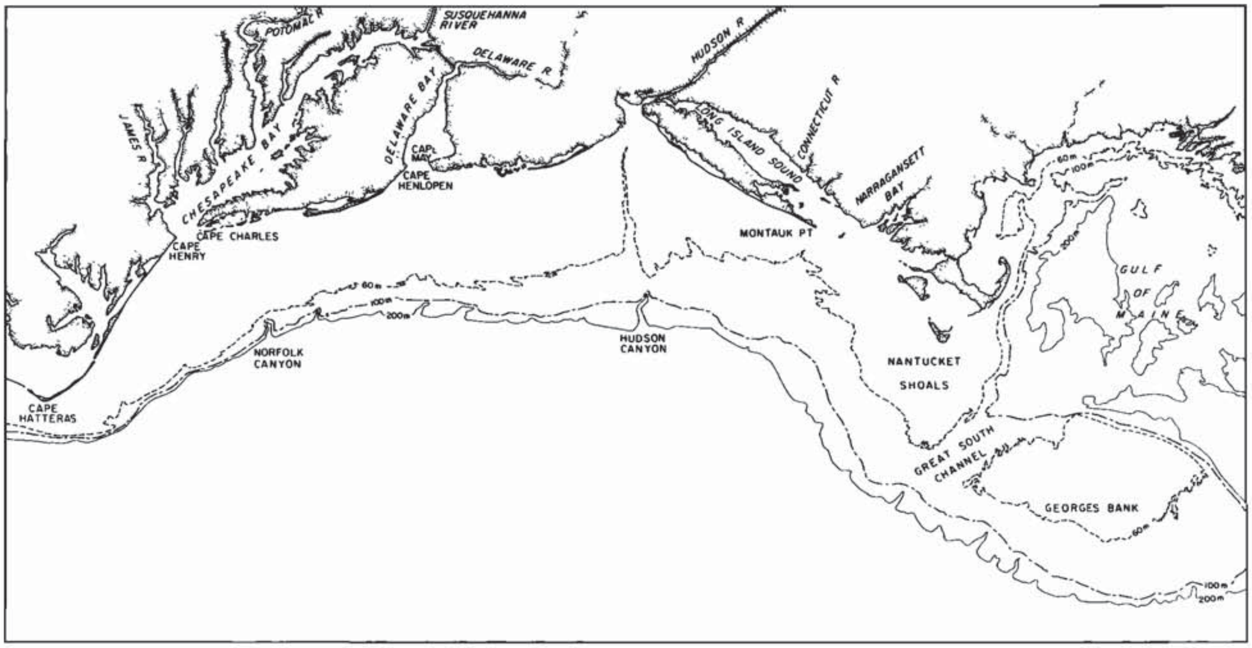

Fig. 13. A topographic map of the Mid-Atlantic Bight, Georges Bank, and a western section of the Gulf of Maine. The 60, 100, and 200 m isobaths are shown [from Beardsley and Boicourt, 1981].

Hydrography

The Mid-Atlantic Bight undergoes a marked seasonal change in stratification [Bigelow, 1933; Bigelow and Sears, 1935; Han and Niedrauer, 1981]. In winter, the water column is often vertically homogeneous over much of the shelf, with the coldest temperatures and freshest water occurring nearshore in February and March (Figure 14). A sharp frontal zone, called the shelf slope front, separates the cooler, fresher shelf water from the warmer, more saline slope water [Wright, 1976]. Originating on the southern flank of Georges Bank, this front runs through the Mid-Atlantic Bight to near Cape Hatteras along the outer shelf and intersects the bottom near the 80 m isobath. In summer, the water column is strongly stratified over most of the shelf due to vernal warming and increased freshwater runoff. The shelf slope front is less distinct in summer, but large horizontal temperature and salinity gradients still exist beneath the seasonal thermocline which has by then developed in both shelf and slope water. The summer stratification is subsequently destroyed in the fall by surface cooling and the seasonal increase in strong synoptic-scale weather disturbances [Mooers et al., 1976].

The shelf slope front and increased mixing nearshore [Pettigrew, 1980] cause the formation in spring and summer of a band of cold, low salinity water called the "cold pool" located near the bottom on the mid and outer shelf (Figure 14). This feature can be traced from the southern flank of Georges Bank to near Cape Hatteras. Houghton et al. [1982] and Ou and Houghton [1982] show that the New England shelf and perhaps Nantucket Shoals is the winter source region for the coldest cold pool water found in the central Mid-Atlantic Bight in the late spring and summer.

Currents

Currents in the Mid-Atlantic Bight can be conceptually divided into tidal currents, subtidal low frequency currents with periods in the 2 to 10 day synoptic band, monthly mean currents, and annual mean currents. Tidal currents are relatively weak in the Mid-Atlantic Bight except in and near some estuaries and sounds and over Nantucket Shoals where tidal currents are sufficiently large to cause complete vertical mixing [Garrett et al., 1978; Bowman and Esaias, 1981; Limeburner and Beardsley, 1982; Ingham, 1982]. Most of the current variability observed in the Mid-Atlantic Bight is associated with winter storms. Storm-driven current fluctuations can be of order 50 cm/s while the observed monthly mean currents over the mid and outer shelf are typically of order 5 to 15 cm/s.

Tidal Currents. Recent tidal current and pressure measurements made in the Gulf of Maine and adjacent shelf regions are summarized by Daifuku [1981], Mayer [1982], and Moody et al. [1983]. These results show that the tidal currents within the Mid-Atlantic Bight are dominated by the rotary semidiurnal components, especially the M_2 component which has a typical maximum amplitude of 10 to 15 cm/s oriented primarily in the cross-shelf direction over much of the region. Redfield [1958] used channel theory to provide the first dynamical model of the M_2 tide south of Long Island. Clarke and Battisti [1981] and Battisti and Clarke [1982a,b] recently showed that the M_2 component can be better modelled within the Mid-Atlantic Bight by

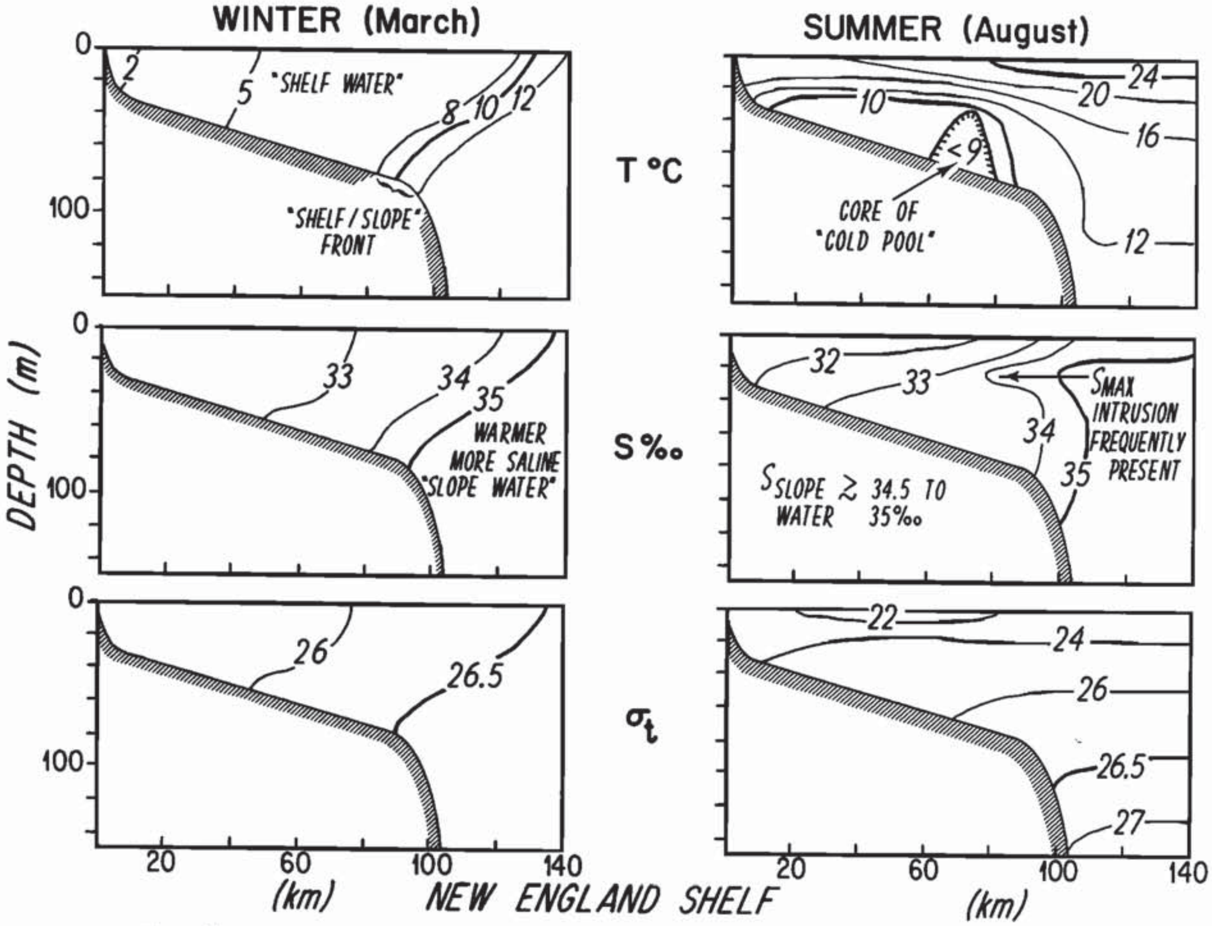

Fig. 14. Schematic cross-shelf temperature, salinity and density (sigma-t) sections for the northern Mid-Atlantic Bight for winter (left) and summer (right).

including the effects of rotation and alongshore sea surface gradients associated with the tides.

The diurnal tidal currents in the Mid-Atlantic Bight are generally weaker than the semi-diurnal components. The K_1 component has a typical maximum amplitude of 2-7 cm/s, oriented primarily in the along-shelf direction, except south of Nantucket Shoals where amplitudes of order 10 cm/s occur associated with a nearby virtual amphidrome. Daifuku [1981] has constructed a barotropic model for the K_1 component containing Kelvin wave, shelf wave, and forced wave components. The Kelvin wave accounts for over half of the surface elevation change while most of the current variance (roughly 80 percent) is associated with the southwestward propagating diurnal shelf wave.

Low-Frequency Currents. General descriptions of the response of the Mid-Atlantic Bight to atmospheric forcing in the 2 to 10 day synoptic band are given in Beardsley and Boicourt [1981] and Csanady [1982]. The synoptic-scale current fluctuations are strongly polarized in the alongshelf direction and account for most (roughly 70 to 90 percent) of the subtidal current variance over the Mid-Atlantic Bight. Over the mid and outer shelf, the alongshelf current and cross-shelf pressure gradient are coherent and in phase, and are approximately in geostrophic balance. Further, both variables are coherent with, but lag, the alongshelf windstress component by roughly 5 to 15 hours [Flagg, 1977; Mayer et al., 1979; Butman et al., 1979; Chuang et al., 1979]. Ou et al. [1981] and Noble et al. [1983] report that the transfer functions over the mid and outer shelf between alongshelf current and windstress components vary with depth from about 10 cm/s per dyne/cm^2 at mid-level to 6 cm/s per dyne/cm^2 at one meter above the bottom. Alongshelf current fluctuations are generally coherent in the cross-shelf plane [Flagg, 1977; Boicourt, 1973; Beardsley et al., 1983], and coherent in the alongshelf direction over separations of order 200 km [Boicourt and Hacker, 1976; Butman et al., 1979; Ou et al., 1981; Noble et al., 1983]. Except near shore in the New York Bight [Pettigrew, 1980; Csanady, 1980, 1982], the alongshelf current is not generally coherent with cross-shelf wind stress.

Recent studies of alongshelf current coherence by Ou et al. [1981] and Noble et al. [1983] support the earlier studies of synoptic-scale coastal sea level variability by Wang [1979] and Noble and Butman [1979] in that the observed low-frequency current fluctuations are dominated by a superposition of a locally wind-driven response and free coastal-trapped waves. Both observations and the model studies of Csanady [1974] and

Beardsley and Haidvogel [1981] suggest that the frictional adjustment time scale for much of the Mid-Atlantic Bight is of order 10 hr, so that the directly wind-driven component should be similar to an arrested topographic wave [Csanady, 1978; Winant, 1979] moving slowly northeastward in phase with the dominant storms. The presence of free coastal-trapped waves in the Mid-Atlantic Bight was recently confirmed by Ou et al. [1981] and Noble et al. [1983] using current measurements. Ou et al. [1981] fitted a two-wave model to current data from two mid-shelf sites separated by 140 km along the 68 m isobath in the New York Bight and found that the forced and free wave components were roughly equal in energy content. The forced wave propagated northeastward while the free wave had a southwestward phase speed of about 525-820 km/day, roughly consistent with an estimate of 600 km/day from the sea level study of Wang [1979]. The presence in this region of equally energetic forced and free waves propagating in opposite directions results in a coherence and rotary phase diagram between untreated velocity records with little indication of alongshelf propagation. Removal of the forced wave component by subtracting that signal coherent with local wind, however, produces clear evidence of southwestward phase propagation [Ou et al., 1981].

While the alongshelf flow is coherent over alongshelf separations of at least 200 km, the low-frequency cross-shelf current fluctuations are incoherent over even the smallest alongshelf separation examined, 70 km [Boicourt and Hacker, 1976; Butman et al., 1979; Mayer, 1982]. Several processes can generate cross-shelf flows with short alongshelf scales. Allen [1976], Chao and Janowitz [1979], Brink [1980], and Wang [1980], for instance, examine the scattering of energy from free continental shelf waves by topographic features such as ridges, canyons, bumps, and regions of isobath convergence. Localized effects associated with bends or embayments in the coastline, or motions generated by small scale spatial wind stress variations can also contribute to the observed alongshelf variability in the cross-shelf flow structure.

A particularly dramatic example of topographic control on the cross-shelf flow has been recently described by Mayer et al [1982]. They find that the Hudson Shelf Valley, which extends almost to the shelf break in the New York Bight (Figure 13) provides a channel for enhanced cross-shelf flow on both synoptic and monthly time scales. Mayer et al. [1982] report monthly mean upvalley flows of 10 to 25 cm/s in winter. In addition to the Hudson Shelf Valley, the series of submarine canyons in the outer shelf provide likely regions for flow disturbance and mixing [Hotchkiss, 1982; Hotchkiss and Wunsch, 1982]. Current fluctuations associated with the shelf slope front [Flagg, 1977; Halliwell and Mooers, 1979] and anticyclonic rings in the slope water [Mooers et al., 1979; Ramp et al., 1983] are additional sources for alongshelf variability in the cross-shelf flow over the outer shelf.

Monthly-Mean Currents. Monthly-mean currents measured from long-term moorings on the outer shelf [Mayer et al., 1979; Beardsley and Boicourt, 1981] support the idea that summertime flow reversals from the generally southwestward mean alongshelf flow occur only in rare years of

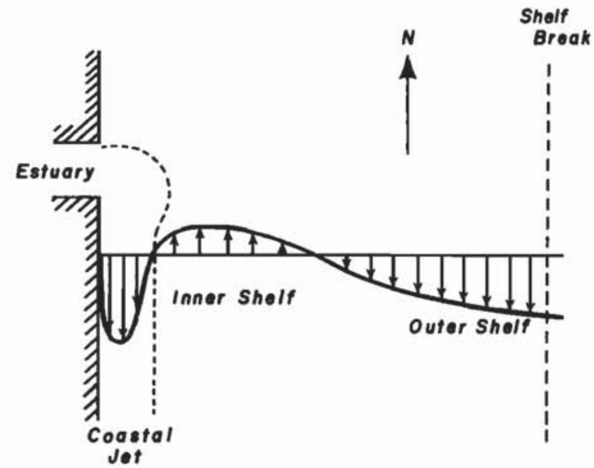

Fig. 15. Schematic alongshore surface velocity profile for summer in the southern Middle Atlantic Bight [from Boicourt, 1982].

strong and persistent northward winds. Bumpus [1969] reported reversals over the inner shelf ("nearshore"), ascribing them to the coincidence of strong northward winds and low runoff from rivers. Recent direct measurements of monthly-mean currents [Beardsley et al., 1976; Boicourt, 1981] suggest that summertime flow reversals over the inner shelf in the southern Mid-Atlantic Bight may be the rule rather than the exception. Current measurements from lightships provide stronger evidence that, while the alongshelf flow over the outer shelf may undergo only occasional reversals, reversals occur annually over the inner shelf. Haight [1942] reports July mean currents (obtained by averaging hourly drift-pole measurements) that are, over a variety of years, consistently reversed with respect to the annual mean. The only exceptions are measurements from lightships at Nantucket Shoals and in estuarine outflow plumes. If near-surface flow over the inner shelf undergoes an annual reversal while the outer shelf reverses only occasionally, then the summertime upper-layer alongshelf flow should be banded into two or three counterflows (shown schematically in Figure 15). The low-salinity estuarine outflows form persistent southward-flowing coastal jets. Beyond this narrow band of width less than 20 km, the inner shelf flow is northward. The width of this band and the shape of the velocity profile offshore of this region are both uncertain. If the profile shown in Figure 15 does represent the July mean nearsurface flow pattern in the southern Mid-Atlantic Bight, then this banded flow structure may exist as a nearly two-dimensional feature from Long Island to Cape Hatteras or as one or a series of mesoscale gyres, with distinct geographical regions of enhanced cross-shelf flow.

In the New York Bight [Mayer et al., 1979] and off Chesapeake Bay [Beardsley and Boicourt, 1981], the measured near surface monthly mean currents are stronger during winter, but the records contain sufficient gaps at this level to preclude defining a seasonal signal. A more consistent seasonal variation is found in the southern Mid-Atlantic Bight where near bottom

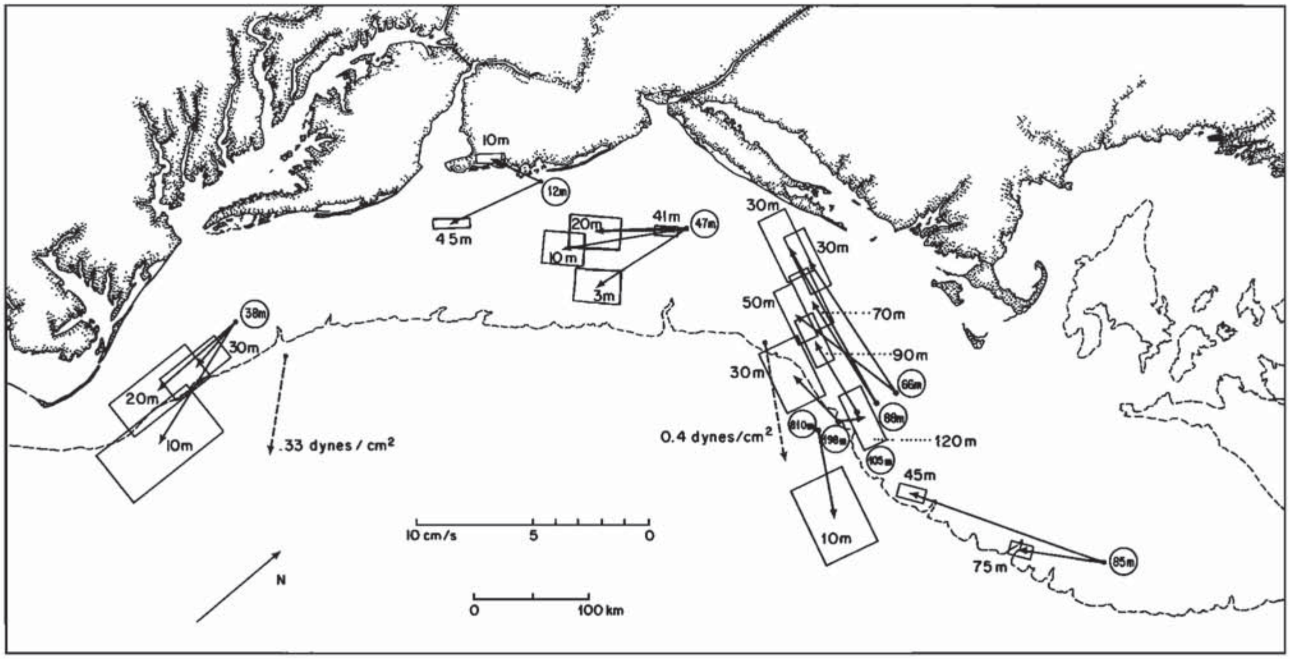

Fig. 16. Map of long-term mean currents computed from one-year or longer moored current observations in the Mid-Atlantic Bight and Georges Bank region. The measurement depth in meters is shown next to the head of the mean current vector, the local water depth in meters at the mooring site is shown next to the tail of the mean current vector. The standard error for each mean current value (estimated using the subtidal current standard deviation, record length, and an assumed five-day correlation time scale) 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 [adapted from Beardsley and Boicourt, 1981 and Beardsley et al., 1983].

currents tend to reverse during one or more winter months while the near surface currents remain southwestward [Figure 7.12 in Beardsley and Boicourt, 1981]. Monthly-mean current data is providing new evidence of seasonal flow variation and differences in the wind-driven component between the northern and southern region of the Mid-Atlantic Bight. There is clearly a need, however, for longer mooring placements (of order 5 years or more) and better spatial coverage and resolution in order to define the seasonal mean circulation and its spatial variation.

Annual Mean Currents. Long-term mean currents computed from one-year or longer direct current measurements clearly demonstrate that the annual mean alongshelf flow is directed towards the southwest throughout the Mid-Atlantic Bight (Figure 16). These mean velocity vectors in water depths shallower than about 60 m show a definite tendency for onshore veering with increasing depth consistent with the drifter results of Bumpus [1973] and the dynamical model results of Csanady [1976]. The immediate upstream source of shelf water in the Mid-Atlantic Bight is the southern flank of Georges Bank and the Gulf of Maine [Hopkins and Garfield, 1981]. Preliminary results using the $H_2^{18}O/H_2^{16}O$ ratio tracer technique suggest that some water within the cold pool in the New York Bight may originate as far upstream as the Gulf of St. Lawrence [Fairbanks, 1982]. Westward advection of shelf water south of Nantucket accounts for approximately 80 percent of the total mean input of fresh water into the Mid-Atlantic Bight.

The dynamical explanation of the observed mean southwestward flow is still unclear. Beardsley and Boicourt [1981] and Csanady [1982] summarize recent work suggesting that the alongshelf flow is driven against an opposing mean wind stress by an alongshelf pressure gradient imposed at the shelfbreak by the cyclonic gyre found between the continental shelf and the Gulf Stream. While a physical coupling between shelf and deep ocean is evident in the numerical model results of Semtner and Mintz [1977], Shaw [1982] and Wang [1982] show that an imposed pressure field cannot penetrate the continental slope in a steady homogeneous ocean, thus baroclinic or nonlinear effects must be important for a basin-scale mechanism to be plausible. A major upstream source of freshwater like the St. Lawrence/Gulf of Maine system may produce a coastally trapped pressure gradient field which extends downstream through the Mid-Atlantic Bight, representing a second possible driving mechanism discussed by Beardsley and Winant [1979], Csanady [1979], Wang [1982] and Shaw [1982].

Cross Shelf Exchange Processes

Iselin's [1939, 1940] inference, from the salinity structure, of mean offshelf flow near-surface and onshelf flow nearbottom has been documented by drifter and moored current measurements [Bumpus, 1973; Beardsley and Boicourt, 1981]. On a synoptic time scale, the cross-shelf current fluctuations reverse with depth during all seasons in a manner consistent with a wind-driven surface Ekman layer [Boicourt and Hacker, 1976; Scott and Csanady, 1976; Flagg, 1977; Mayer et al., 1979; Chuang et al., 1979]. The low alongshelf coherence of cross-shelf flow and the possibility of regional circulation cells and of locally enhanced cross-shelf flows due to subtle topographic effects have generally frustrated attempts to measure and describe the cross-shelf transport of salt and heat [Bush, 1981; Fischer, 1980]. Two-dimensional simplifications are clearly inadequate for both observational schemes and analytic approaches attempting to construct accurate regional heat and salt budgets.

Most observations of exchange between shelf water and slope water have consisted of temperature and salinity measurements from shipboard. For this reason, the resulting picture is primarily a catalog of processes on an event or synoptic time scale, with temporal or spatial coverage that seldom allows a quantitative assessment of the associated transports. These processes include: "calving," the formation of discrete parcels of shelf water at the shelf slope front [Cresswell, 1967; Wright, 1976; Wright and Parker, 1976]; entrainment of shelf water in the trailing edge of anticylconic warm core rings [Saunders, 1971; Morgan and Bishop, 1977; Scarlet and Flagg, 1979; Smith, P. C., 1978]; cabelling [Garrett and Horne, 1978; Bowman and Okubo, 1978]; interleaving [Voorhis et al., 1976; Houghton and Marra, 1982]; and intrusions of high salinity slope water onto the Mid-Atlantic Bight which can occur in both stratified and unstratified seasons (see Figure 14). Boicourt and Hacker [1976] report that large intrusions extending more than 50 km across the shelf in a layer 5 to 10 m thick during summer in the southern Mid-Atlantic Bight are correlated with strong northeastward winds. They interpret these intrusions to be the onshelf flow at mid-depth above the cold pool in compensation for a wind-driven off-shelf flow in the upper Ekman layer. The regular occurrence of smaller intrusions over the outer shelf, even in the absence of strong winds, has led Posmentier and Houghton [1980], Gordon and Aikman [1981], and Welch [1982] to offer other mechanisms for their generation.

Until observational coverage and resolution suffice to allow a reliable, direct estimate of the cross-frontal and cross-shelf fluxes associated with these processes, the alternate approach of estimating salt and heat fluxes from a model of the observed seasonal progression over the entire Middle Atlantic Bight may prove more accurate. Ou and Houghton's [1982] model incorporating Houghton et al.'s [1982] observations of the cold pool temperature is a successful example of such an approach. A clear picture of the relative importance of the individual cross-shelf exchange processes contributing to the salt and heat budgets, however, has yet to be developed.

Georges Bank
(B. Butman)

Georges Bank is a large shallow submarine bank located along the southeastern side of the Gulf of Maine (Figure 13). The bank is about 300 km long and 150 km wide. The water on the crest of the bank is 30-40 m deep although less than 5 m deep in some places. The bank is separated from the Scotian Shelf by the Northeast Channel (220 m deep) and from Nantucket Shoals by the Great South Channel (70 m deep).

Studies of the seasonal mean circulation in the Georges Bank-Gulf of Maine region using surface and bottom drifters were conducted by Bigelow [1927] and Bumpus [1973, 1976]. These studies showed a residual surface counterclockwise circulation in the Gulf of Maine and a clockwise circulation around Georges Bank. The clockwise circulation around Georges Bank was strongest in spring and summer. In winter, the near-surface flow was primarily offshore, driven by the northwesterly winds.

Beginning in 1975, extensive current and hydrographic measurements were made on Georges Bank and the adjacent shelf and slope. Selected results from the recent field programs and a description of some of the modelling efforts are presented here. Results of recent studies conducted on the adjacent Scotian Shelf may be found in Smith [1983] and of observations made in the Northeast Channel in Ramp et al. [1983b].

Hydrography

The water on the crest of Georges Bank (depths shallower than 60 m) is vertically well mixed throughout the year by the strong semidiurnal tidal currents [Colton et al., 1968; Bumpus, 1976; Garrett et al., 1978]. The well-mixed Georges Bank water is warmest, freshest, and lightest (14°C, 32.5‰ and 24.2 sigma-t) in summer, and coldest, saltiest, and heaviest (5°C, 33.3‰ and 26.4 sigma-t) in winter [Hopkins and Garfield, 1979; Flagg et al., 1982]. In winter, two fronts separate the well-mixed water from adjacent water masses (Figure 17a-c). On the southern flank of the bank, the shelf-water/slope-water front intersects the bottom at approximately 80 m and separates cooler, fresher shelf water from warmer, more saline slope water. The shelf-water/slope-water front is similar in structure and continuous with the front at the shelf break in the Middle Atlantic Bight [see Beardsley and Flagg, 1976; Wright, 1976; Mooers et al., 1979]. On the northern flank, a second weaker and deeper front separates Georges Bank water from Gulf of Maine water. In summer (Figure 17d-f), a seasonal thermocline develops over the Gulf of Maine, the slope water, and the water deeper than 60 m on the southern flank. A tidally mixed front forms at approximately the 60-m isobath. A subsurface bank of cool water, referred to as the "cold pool," occurs along the southern flank of the bank between the 60-m and 100-m isobaths, bounded by the warmer slope water to the south, the warmer well-mixed Georges Bank water to the north, and the seasonal thermocline above. Houghton et al. [1982] have described in detail the structure and changes of the cold pool on the southern side of Georges Bank and in the Middle

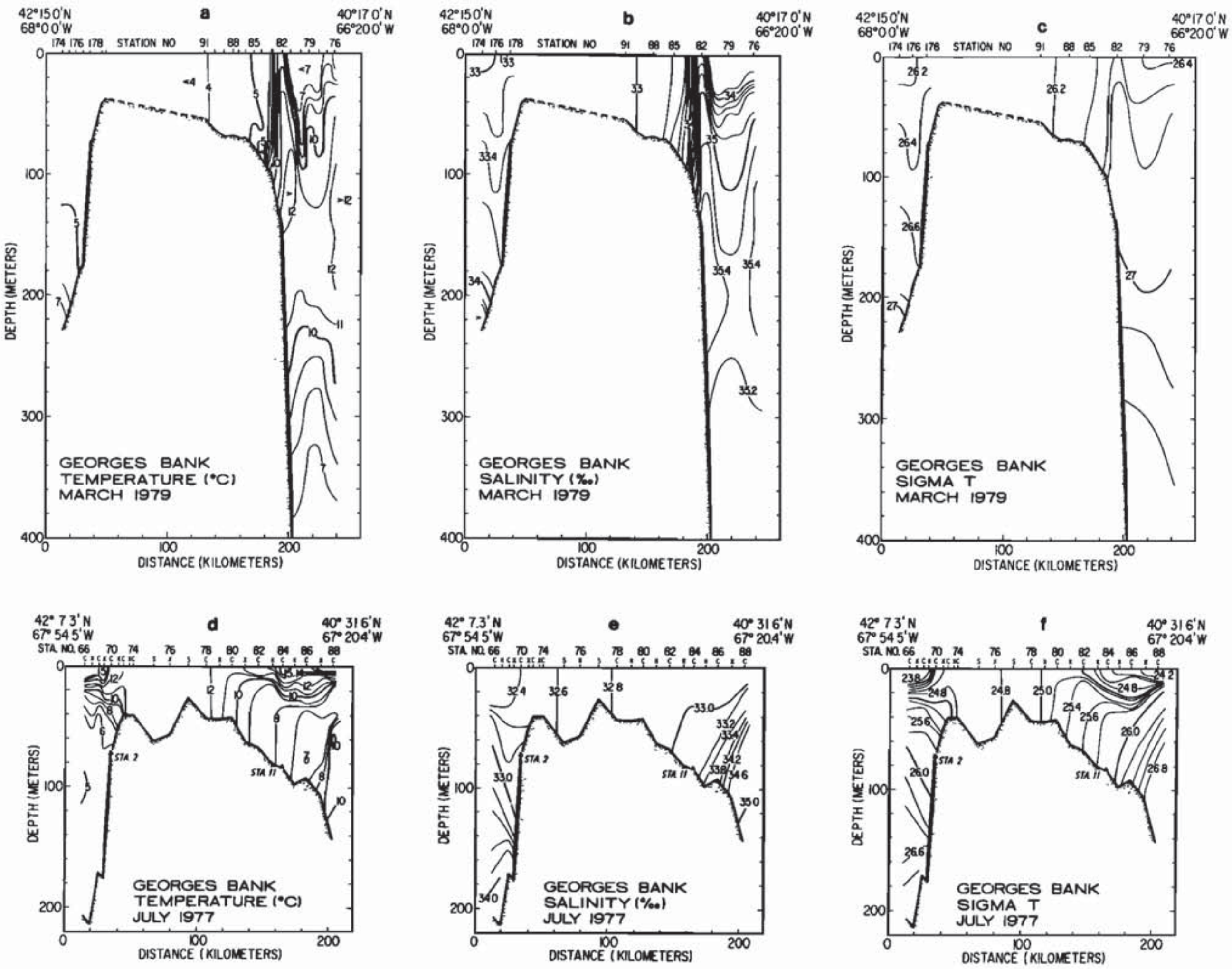

Fig. 17. Typical hydrographic sections running roughly from northwest to southeast across the crest of Georges Bank. (a)-(c): Winter sections showing well-mixed water on the crest and on the northern and southern flanks to water depths of 80-100 m. These sections were taken in March 1979 at the winter temperature minimum. (d)-(f): Summer sections showing well-mixed water on the crest, tidal fronts at ∿40 m on the northern flank and ∿60 m on the southern flank, the cold band on the southern flank and the shelf-water/slope-water front which intersects the shelf at ∿80 m. These sections were taken on July 8-9, 1977. X indicates stations with temperature observations only, C indicates stations with temperature, salinity, and density observations [from Butman et al., 1982].

Atlantic Bight during 1979. Within the well-mixed region, the concentrations of nutrients (nitrate, phosphate, and silicate) and of oxygen are highest in winter and lowest in mid-summer [Pastuszak et al., 1982]. Loder et al. [1982] used a simple axisymmetric model of the evolution of temperature in spring and summer to study horizontal exchange on central Georges Bank. The analysis suggests horizontal diffusion coefficients of order 150-380 m^2/s.

Currents

Conceptually, the currents can be divided into a seasonal mean circulation, subtidal low-frequency currents with periods of 50-600 hours, and energetic tidal currents. Variance conserving kinetic energy spectra (Figure 18) of the currents at 45 m at station 11 on the southern flank of Georges Bank (see Figure 19 for station location) clearly show that the semidiurnal and diurnal tidal currents dominate the currents on Georges Bank. At this station, the amplitudes of the semidiurnal (M_2) and diurnal tidal currents (K_1) were 38 and 5 cm/s, respectively. The amplitude of the subtidal currents were typically 5-15 cm/s, and were fairly evenly distributed over periods from 50 to 600 hours. Note that in the subtidal band, the maximum in the wind-stress spectra occurred at higher frequencies than the current spectra maximum. This suggests that not all of the subtidal currents at this location were

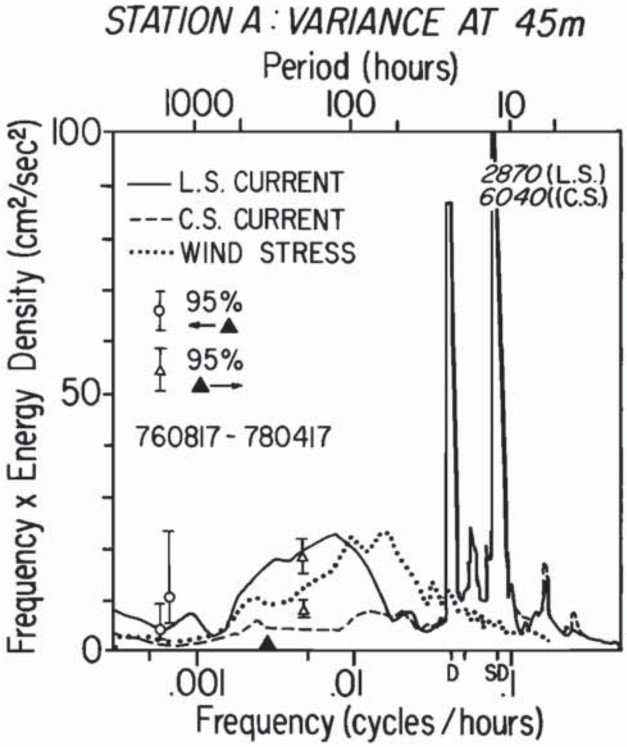

Fig. 18. Variance conserving spectral plot of currents at 45 m at station 11 (water depth 85 m, see Figure 19 for station location) and wind stress at Nantucket Lightship for the period August 1976 to April 1978. The amplitudes of the energetic (and off-scale) along-bank and cross-bank semidiurnal tidal currents are indicated in parentheses. Confidence limits are shown as open triangles for the region to the right of the solid triangles and as open circles for the region to the left [from Butman and Beardsley, 1982].

directly wind driven. The mean current and current fluctuations with periods larger than about 140 days are not shown by the energy spectra. However, the mean current was about 8 cm/s, and there was a seasonal variation of the monthly mean current of about 3 cm/s at this location. Thus, at station 11, the amplitude of the mean along-bank flow, the seasonal variation of the along-bank flow, and the subtidal currents were approximately the same amplitude. Note that the cross-bank currents were generally weak except at tidal periods.

Seasonal Mean Circulation. The recent Eulerian subsurface current measurements made in the Georges Bank region indicate that a residual subsurface clockwise circulation exists around Georges Bank throughout the year [Butman et al., 1982]. The circulation (Figure 19) is defined by a broad southwestward flow along the southern flank, a northward flow on the eastern side of the Great South Channel into the Gulf of Maine, a strong northeastward flow along the northern flank, and a southerly flow on the northeast peak. Current speeds were typically 5-10 cm/s; thus water particles could circuit Georges Bank in about 2 months. The circulation apparently diverged south of Great South Channel where some flow turned northward into the Gulf of Maine and some continued westward into the Middle Atlantic Bight. Observations of drifters drogued at 10 m suggest that water particles which passed south of Great South Channel were most likely to turn northward into the channel in summer and least likely in winter [Flagg et al., 1982]. Water from the southern flank of Georges Bank may reach the area south of Cape Cod in the Middle Atlantic Bight in about 30 days. The circulation pattern determined from these recent direct current observations is similar to the summertime surface circulation described by Bigelow [1927] and Bumpus [1973]. However, the recent observations show that the clockwise circulation exists throughout the year and extends to the subsurface currents. Substantial seasonal and low-frequency variability in the along-bank flow was observed, however, and thus the flow pattern often varies from the simple clockwise circulation.

Nearly continuous current measurements at 45 and 75 m were made at station 11 on the southern flank of Georges Bank from May 1975 to March 1979 [Butman and Beardsley, 1982]. These long-term current observations show that although the monthly averaged flow was southwestward throughout the year, the strength of the along-bank flow ranged from -2 to -17 cm/s and varied seasonally. On average, maximum along-bank flow at 15 and 45 m was in August-September, and minimum flow was in March (Figure 20). There was no significant seasonal variation in the along-bank flow at 75 m (10 m above bottom). Shorter term current observations at other stations around the periphery of Georges Bank suggest a similar seasonal pattern [Flagg et al., 1982; Butman and Beardsley, 1982].

Rectification of the strong semidiurnal tidal curents which flow across Georges Bank into the Gulf of Maine may partially drive the observed around-bank flow [Loder, 1980; Hopkins and Garfield, 1981]. Along-bank currents predicted by Loder's depth-averaged analytical model occurred in relatively narrow regions over the steeply sloping topography on the sides of the bank. The predicted current amplitudes were about 10-25 cm/s on the northern flank and 6 cm/s on the broader southern flank. These amplitudes are of the same order of magnitude as the observed mean flow. An analysis of the current observations for evidence of tidal rectification suggests that only a portion of the observed mean along-bank flow may be caused by tidal rectification, however. Butman et al. [1983] used realistic topography for the southern flank of the bank and extended Loder's model to include the weaker N_2 and S_2 tidal components. The analytic results show a weak jet of about 3 cm/s at the shelf break, a relative minimum (less than 1 cm/s) on the southern flank, and increased tidally rectified along-bank flow toward the crest of the bank. Butman et al. [1983] used the ratio of the modulation of the along-bank flow at fortnightly and monthly periods to the mean along-bank flow predicted from the analytical model, and the observed (weak) modulation to estimate the mean current caused by tidal rectification at station 11. The model and observations indicate that <u>less</u> than about 40% of the mean along-bank flow at station 11 was caused by tidal rectification (about 2 cm/s). Magnell et al. [1980] computed

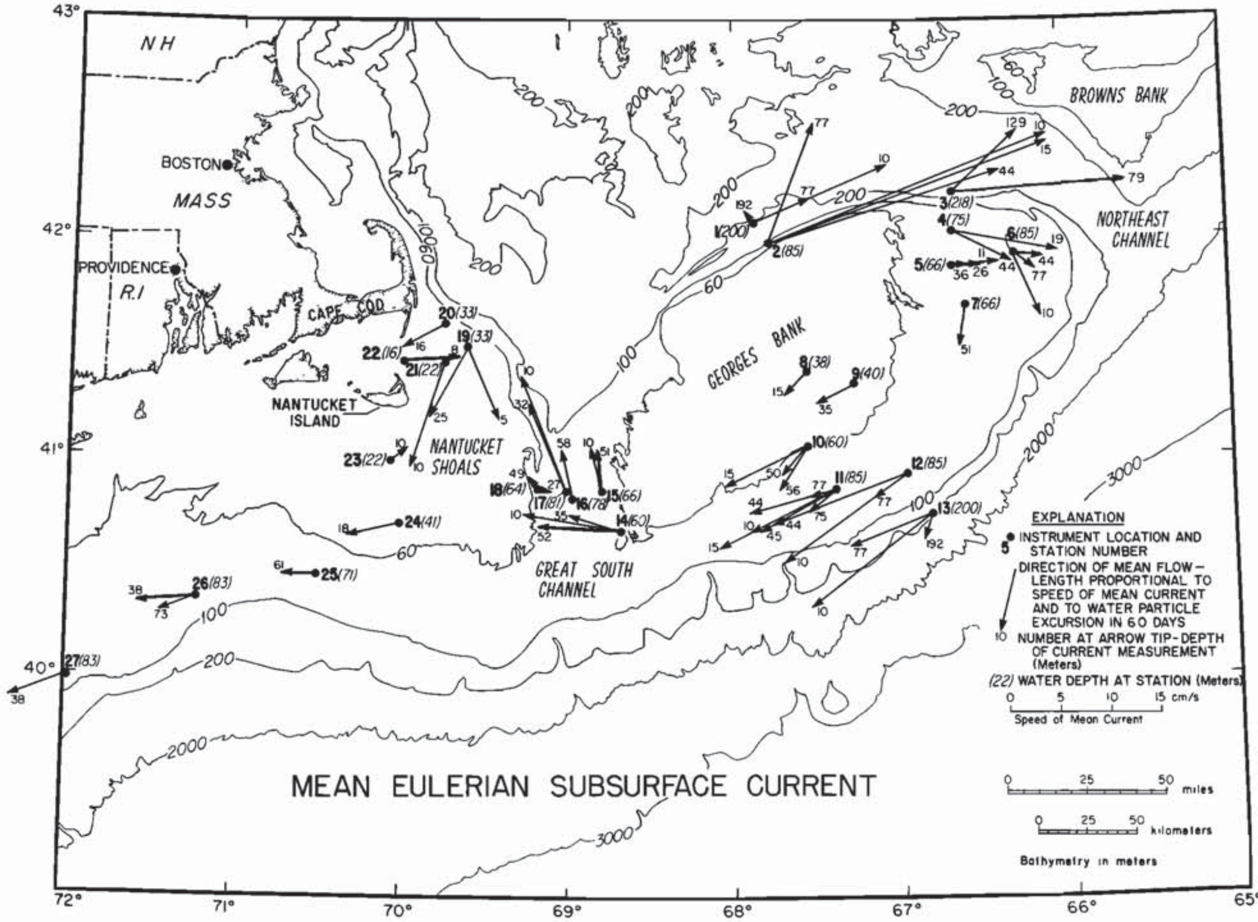

Fig. 19. Mean Eulerian current measurements. The boldface number at origin of vector is the station identifier. The number in parentheses following identifier indicates water depth at that station. The number at the tip of the vector indicates the depth of measurement in meters. The length of vector is proportional to the mean current speed. The speed scale is such that the length of the current vector is equivalent to the mean displacement of a water particle during a 6-day period at map scale or a 30-day excurison at five times map scale. The current measurements were not synoptic nor of equal duration but are compiled from measurements of varying length (1 month minimum) made at different times during the 4-year period, 1975-79 [from Butman et al., 1982].

the coherence between the amplitude of the observed cross-bank tidal current and the along-bank current at one location on the steeply sloping northern flank where the amplitude of the tidally rectified current should be large. The coherences were low. However, due to the expected temporal and spatial variability of the narrow jet, Eulerian current measurements may be difficult to interpret in this region. Although these preliminary analyses suggest that not all of the observed along-bank flow is caused by tidal rectification, they are single-point measurements and are not inconsistent with the existence of strong tidally rectified currents in other regions of the bank. Detailed numerical models may be the best way to determine the importance of tidal rectification on Georges Bank [Greenburg, 1983].

Cross-bank and along-bank density gradients may also drive part of the around-bank flow. Estimates of the cross-bank density field show lighter water on the crest of the bank in summer, consistent with the stronger observed around-bank flow. Cross-bank density gradients were strongest across the northern flank of the bank in the region of the swift northeastward jet [Flagg et al., 1982]. Butman and Beardsley [1982] found the observed seasonal shear between 45 and 75 m at station 11 was explained entirely by the observed cross-bank density field, but the seasonal shear between 15 and 45 m was not explained by the available density observations. It was also not clear if the increase in the southwestward flow at station 11 in summer indicated an increase in the around-bank flow. The shelf-water/slope-water front was onshore at the same time as the maximum along-bank flow; thus the increased southwestward flow at station 11 could be interpreted as a constant southwestward transport through a decreased cross-sectional area inside of the front. The increase in the gradient of the cross-bank density field at shallower stations near the top of the bank was consistent with an increased around-bank flow, however.

These data suggest that the seasonal mean circulation around Georges Bank is composed of a steady component which exists throughout the year, probably caused by tidal rectification, and a seasonally varying component partially caused by the cross-bank density field. This conceptual model will be refined with additional analyses and analytical and numerical models. The contribution of wind stress, an along-bank pressure gradient, and along-bank density gradients to the residual circulation remains to be determined, perhaps using simple diagnostic models [Csanady, 1974, 1976; Smith, 1983].

Low-Frequency Currents. The amplitude of the low-frequency currents in the Georges Bank region are typically 5-15 cm/s [Butman et al., 1982; Flagg et al., 1982]. The low-frequency currents were aligned parallel to the local isobaths and were stronger near the surface than near the bottom.

The major portion of the low-frequency along-bank current variance was concentrated in the wind-driven frequency band of 2-10 days (Figure 18). However, the coherence between wind stress and currents indicates that typically less than 50% of the low-frequency current variance

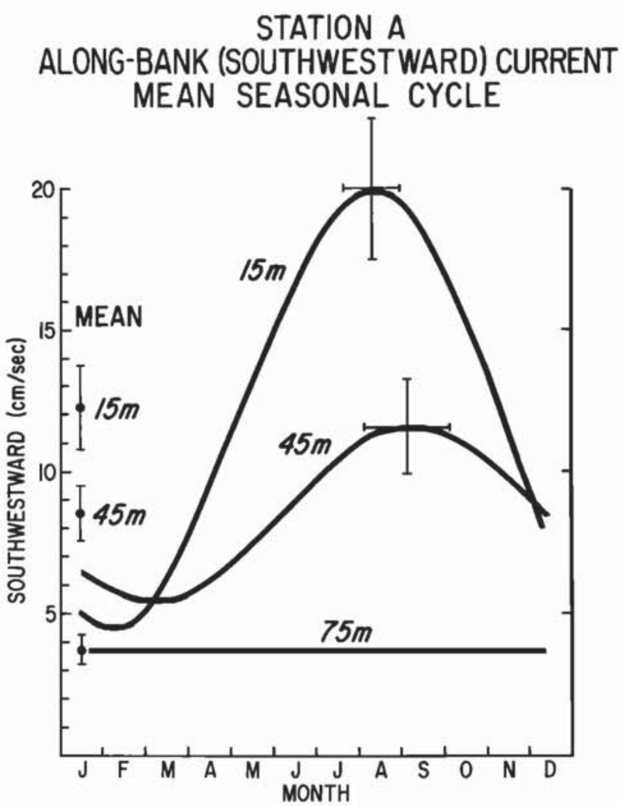

Fig. 20. Seasonal cycle of along-bank current at station 11 determined by least squares fit to the monthly averaged data. Error bars are the 95% confidence limits. Positive flow is toward the southwest.

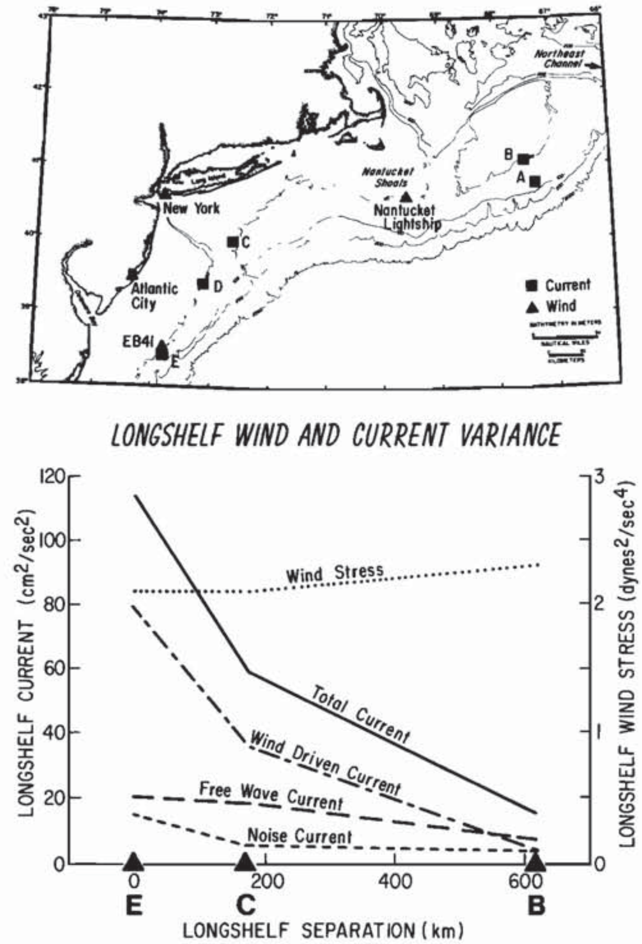

Fig. 21. Location of near-bottom current measurements and wind observations (top) and the partition of near-bottom long-shelf low-frequency current variance between wind-driven, free-wave, and noise components as a function of long-shelf position for data obtained at stations B, C, and E from January 15-March 15, 1978 (bottom). The variance has been averaged over the periods from 60 to 360 hours. Wind stress at station B is from Nantucket Lightship, wind stress at stations C and E was determined from JFK International Airport, increased by the average winter ratio of winds observed at EB41 and JFK International Airport in winter [from Noble et al., 1983].

was driven by the wind [Flagg et al., 1982; Noble et al., 1983]. This is in contrast to the shelf southwest of Georges Bank where the along-shelf currents were more correlated with wind stress. Mayer [1982] found that greater than 60% of the low-frequency current variance was correlated with wind stress in the Middle Atlantic Bight.

In a comparison between low-frequency currents observed on Georges Bank and in the Middle Atlantic Bight (MAB) in winter, Noble et al. [1983] separated currents measured 1 m above the bottom at three locations along the 60-m isobath into wind-driven currents and currents associated with freely propagating waves; 70-90% of the energy in the long-shelf currents in winter were attributed to these two components (Figure 21).

The amplitude of the near-bottom subtidal longshelf current was much larger in the MAB than on the southern flank of Georges Bank (about 10-15 cm/s and 5 cm/s, respectively). The wind-driven currents accounted for the major portion of the increased current variance in the MAB; wind-driven currents were 8-12 cm/s in the MAB and only about 3 cm/s on the southern flank of Georges Bank. The wind stress was relatively uniform over the MAB-Georges Bank region and thus the larger wind-driven currents in the MAB were not caused by stronger winds. The free-wave currents also increased southwestward along the shelf from about 4 cm/s on Georges Bank to 6 cm/s in the MAB. Thus, the subtidal currents on the southern flank of Georges Bank were weaker than in the MAB, and free waves and wind-driven currents were of approximately equal importance. Ou et al. [1982] also separated currents measured on Georges Bank and in the MAB into southwestward propagating free waves and a wind-driven component.

These observations show that the low-frequency current variability in the Georges Bank region is complex and only partially driven by wind stress, and that there is significant longshelf structure to the subtidal current variability. The absence of a simple coastal barrier across the top of the bank probably complicates the flow field [Brink, 1983]. Noble et al. [1983] have suggested that the along-shelf structure of the low-frequency currents may be partially related to the scale of the coastal wind systems [Csanady, 1978]. Gulf Stream eddies impinge along the outer edge of the shelf and also affect currents in the low-frequency band [Flagg et al., 1982; Friedlander, 1982]. Further analysis is required to describe and understand the low-frequency current dynamics of the Georges Bank region. An understanding of the entire Georges Bank-Gulf of Maine-Middle Atlantic Bight-Scotian Shelf system may be required to fully determine the low-frequency current dynamics.

Tidal Currents. The currents in the Georges Bank region are dominated by strong rotary semidiurnal tidal currents. Bumpus [1976] summarized the available tidal current data, primarily reported by Haight [1942]. Greenberg [1979] developed a numerical model of the tide in the Georges Bank-Gulf of Maine region. Tidal observations made as part of the recent field programs are summarized by Moody et al. [1983].

The tidal currents are strongest (50-100 cm/s) on the shallow crest of the bank and on Nantucket Shoals, weaker on the northern and southern flanks, and extremely weak (less than 5 cm/s) south of Cape Cod. The strong tidal currents determine several important characteristics of the region. As discussed previously, rectification of the semidiurnal tides may partially drive the along-bank flow. The tidal currents also partially determine the hydrography through vertical mixing. The tidal currents are strong enough to vertically well mix the water column throughout the year on the crest of the bank, in the Great South Channel, and on Nantucket Shoals [Garrett et al., 1978]. A seasonal thermocline forms in water depths deeper than about 60 m, where the tidal currents are too weak to keep the water column well mixed. The tidal currents partially control the texture of the surface sediments and the surface topography. Coarse sand and large mobile bed forms are found on the crest of the bank where the tidal currents constantly rework the surficial sediments [Twichell et al., 1981; Twichell, 1983]. On the flanks of the bank where tidal currents are weaker, the sediments are finer and the sea floor is featureless. The tidal currents are weakest in the region of fine-grained sediments south of Cape Cod. Twichell et al. [1981] and Bothner et al. [1981] suggest that this modern deposit is composed of fine sediments winnowed from the crest of the bank by storms and tidal currents and carried westward in the mean flow. The sediments are deposited south of Cape Cod where the tidal currents decrease and allow the fine sediments to drop out of suspension.

Peru
(J. S. Allen)

The coastal region off Peru is well known as a location of persistent coastal upwelling and of high biological productivity. The Coastal Upwelling Ecosystems Analysis (CUEA) program conducted a large joint physical and biological field experiment, called JOINT-II, off Peru near 15S from March 1976 to May 1977. Intensive experiments were conducted in March-July 1976 and March-May 1977 (MAM 77). During these time periods, currents and temperatures were measured from several moorings deployed on the continental shelf and upper slope near 15S (Figure 22) and also from single moorings spaced alongshore equatorward to 10S. Wind observations were obtained from shore stations and, in 1977, from buoys. Sea surface temperature and near surface wind fields were mapped by aircraft, and an extensive hydrographic program was implemented. In MAM 77, the latitudinal extent of the current measurements was effectively extended by observations made at 5S in a joint German-Peruvian experiment [Brockmann et al., 1980]. Information on the seasonal variability of the currents at mid-shelf was obtained by maintaining the mooring M (Figure 22) for the entire 14 month experiment.

Geographically, the Peru coast provides an interesting contrast to the regions discussed in other sections. The shelf is on an eastern ocean boundary, at low latitudes in close proximity to the equator. Unlike mid-latitude shelves, the internal Rossby radius at 15S is greater than the shelf width. In addition, the equatorial waveguide provides a theoretically possible additional source region for fluctuating disturbances. Both of these features appear to play a large role in determining the nature of the observed flow.

Winds

The large-scale, low frequency (periods >2 days) winds along the coast of Peru are associated with the southeast trades and are extremely persistent in both direction and magnitude [Enfield, 1981a,b]. The dominant direction is alongshore and equatorward and hence upwelling favorable. The magnitudes, for example, of the mean and standard deviation of the alongshore component of the wind stress τ, measured from a nearshore buoy at 15S in MAM 77, were 0.8 and 0.4 dynes/cm^2, respectively [Allen and Smith,

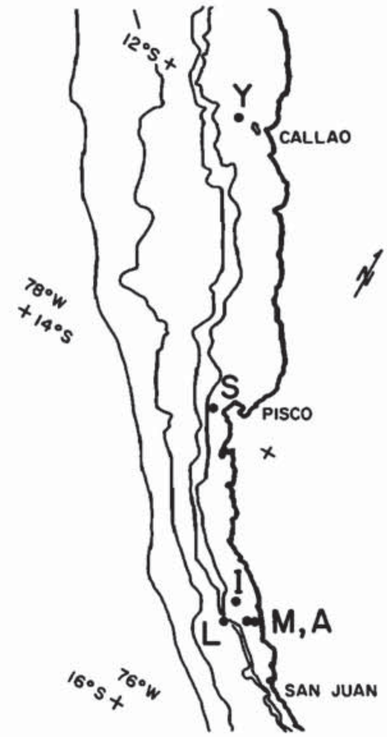

Fig. 22. Peru continental shelf and slope between 12S and 16S. Isobaths are 100, 200, 1000, and 2000 fathoms. The locations of the JOINT-II experiment current meter moorings for March-July 1976 are shown.

1981]. In March-April 1976, the magnitude of the monthly mean alongshore wind stress measured on the coast at San Juan was similar to the MAM 77 value. It subsequently increased steadily to about 1 dyne/cm^2 in August-September 1976 [Brink et al., 1978] consistent with the annual cycle discussed by Enfield [1981b]. An example of winds measured at San Juan in 1976 are shown in Figure 23.

Considerable energy also exists in the wind field on short time and space scales. Buoy measured winds in MAM 77 exhibited significant peaks in the wind stress spectra at the diurnal period, evidently reflecting a strong thermally induced sea-breeze circulation [Brink et al., 1980]. Appreciable spatial gradients in magnitude were also evident. In the cross-shelf direction, larger mean wind speeds were measured near shore. In the alongshore direction, larger mean speeds were found near the main current meter moorings, with decreasing values north and south. This general spatial pattern was also found in mean streamline and isotach maps of wind velocities measured at 152 m height from aircraft during 12 flights in MAM 77 [Moody et al., 1981; Stuart, 1981; Stuart et al., 1981]. The alongshore variation was striking, with a decrease in magnitude from roughly 12 m/s near the moorings to 4 m/s at a distance 30 km to the north. This spatial pattern was attributed by Moody et al. [1981] to a thermally induced circulation associated with the Ica River Valley which penetrates the coastal mountain range at about 14.9S.

Currents and Hydrography

The most extensive set of measurements was made in MAM 77. In particular, surface layer velocities were measured at two locations on the shelf (near M and A, Figure 22). Combined with data from nearby subsurface moorings, this effectively resulted in velocity measurements throughout the water column at two locations [Brink et al., 1980]. Consequently, the observations referred to here will generally be from MAM 77. Similar or consistent results were obtained from the 1976 observations, except where noted. It will be convenient to describe the flow in terms of frequency ranges with the lowest range here provided by time averages over the three month intensive experiments, such as MAM 77, or similar periods in 1976. These will be referred to as seasonal mean values. Low frequency will then designate fluctuations about the seasonal mean values with periods greater than about 2 days. Inertial, diurnal, and semi-diurnal period motions will be termed intermediate frequency.

The bottom topography of the study region is shown in Figure 22. At 15S, the shelf is narrow

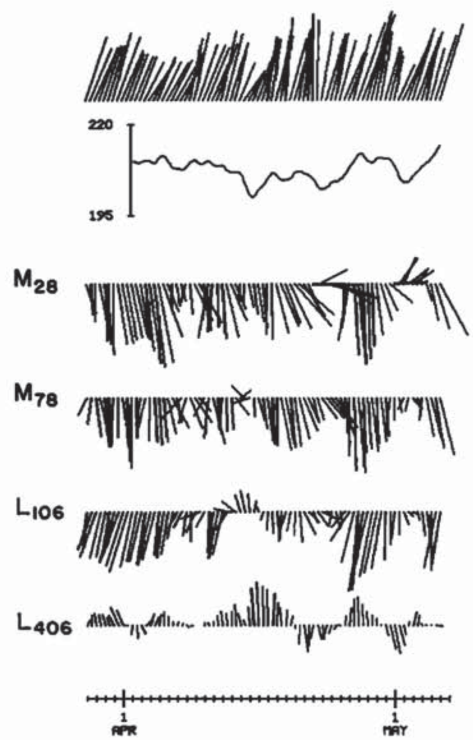

Fig. 23. Wind velocities (top) and coastal sea level, both measured at San Juan, and current velocities measured at moorings M and L during April 1976. All records have been low-pass filtered (half power point 0.6 cpd). Water depths at M and L are 123 and 656 m, respectively. Depths in meters of current measurements are given by subscripts. Scale: 35 cm on sea level scale corresponds to 10 m/s for winds and 50 cm/s for currents. Winds are rotated 45° toward northwest (alongshore) and currents are rotated into a mooring average principal axes system: M (39°) and L (62°) [from Brink et al., 1978].

(offshore distance to the shelf break at 150 m depth is ∿20 km) and the slope is steep. The width of the shelf varies along the coast, being greater equatorward of 14S.

Seasonal Mean. The cross-shelf structures of the mean current and the mean temperature fields at 15S during MAM 77 are given in Brink et al. [1980]. The mean alongshore velocity v is equatorward in the direction of the mean alongshore wind stress only in a shallow layer of depth about 20 m and at depths below 200 m. The remainder of the mean alongshore flow over the shelf and upper slope is poleward, opposite in direction to the mean wind stress. The maximum magnitudes are typically 15 cm/s equatorward in the surface layer and 6 cm/s poleward at 60 to 100 m depth. The mean poleward flow is referred to as the poleward undercurrent and is evident in the vector plots of current velocities in Figure 23.

The mean temperature field has a cross-shelf structure that is consistent with a thermal wind balance of the mean alongshore velocity field. The isotherms above 50 m depth slope upward toward the coast, reflecting the effects of coastal upwelling, while those below 100 m depth slope downward toward the coast.

The current measurements, which extend offshore about 25 km, do not indicate an offshore decay of the mean alongshore velocities. Offshore scales were found by Huyer [1980], who computed alongshore geostrophic velocities at various offshore positions, out to 100 km from the coast, using a series of 35 repeated hydrographic sections. Over the upper slope, the mean vertical profiles of the computed geostrophic velocities were in reasonable agreement with the mean values from current meter measurements. Huyer [1980] concluded that the poleward undercurrent has largest velocities at about 100 m depth near the slope and that it is confined primarily within an offshore scale of 60 km.

The seasonal mean poleward undercurrent was generally stronger during the same period in 1976 [Brink et al., 1978], a fact that may be related to the presence of El Niño conditions during that year. It was present at mid-shelf during the total 14 month experiment [Romea and Smith, 1983] and therefore appears to exist during all seasons. Information on the alongshore scale of the undercurrent was obtained from simultaneous current measurements at 5, 10, 12, and 15S during part of MAM 77 [Brockmann et al., 1980]. The undercurrent was found at all locations and thus, at that time, extended from at least 5 to 15S. After scaling the offshore distance by the internal Rossby radius, the horizontal and vertical structure at 5 and 15S were found to be similar, with the magnitudes of the mean alongshore currents larger at 5 than at 15S.

The mean cross-shelf velocities at 15S are offshore in the upper 20 m, as expected from Ekman theory with an equatorward mean τ. Over the mid- and inner shelf (depths $\leq$120 m), the mean velocities are onshore below 20 m, with weak offshore flow within 10-20 m of the bottom [Brink et al., 1980]. The mean mid-shelf onshore and offshore transports do not balance. The offshore transport in the upper layer is smaller than the mean onshore transport in the lower layer, which in turn is close in magnitude to the mean Ekman transport computed from τ [Smith, R. L., 1981]. The nonzero net cross-shelf transport implies that alongshore gradients are important at this location and that the mean flow is three-dimensional.

The mean sea surface temperature maps [Moody et al., 1981] also show marked alongshore variations. There is a cold spot roughly centered in the location of the current meter moorings. The coldest water (T < 17C) is next to the coast. Temperatures generally increase offshore, but there are also large gradients north and south along the coast. A relative warm spot (T > 19.5C) exists about 40 km north. The scales and alongshore patterns are similar to those found in the maps of the mean wind field, with colder water corresponding approximately to larger wind velocities. Subsurface temperature and hydrographic measurements indicate that these surface temperature gradients are primarily confined to shallow water, above 50 m depth [Brink et al., 1980; Brink et al., 1981].

Low Frequency. The alongshore velocity fluctuations are well correlated with each other on the shelf and over the upper slope to depths of about 250 m [Brink et al., 1978; Brink et al., 1980]. High correlations of v with coastal sea level and with cross-shelf bottom pressure differences and high correlations of vertical gradients of v with cross-shelf gradients of temperature indicate that the low frequency alongshore velocity is in geostrophic balance, outside of surface and bottom boundary layers with thicknesses of order 10-20 m [Brink et al., 1980]. On the shelf, the interior fluctuations in v have relatively weak vertical shear, while over the slope there is considerable shear between, e.g., 200 and 500 m. These features are evident in the vector plots in Figure 23. The first mode of the "section" empirical orthogonal functions (EOF's) for v, calculated using all current measurements from the main cross-shelf line of moorings, accounts for 54% of the variance and describes the major part of the time-dependent v field on the shelf and upper slope [Brink et al., 1980]. Unlike other upwelling regions, the fluctuations in the alongshore currents outside of the surface layer are not correlated with the local wind stress.

The cross-shelf velocities and the temperature show more evidence of direct wind forcing. At mid-shelf, fluctuations in the depth-integrated cross-shelf transport of the surface layer and of the lower layer beneath that are significantly correlated with the alongshore component of the wind stress in a sense consistent with Ekman transport at the surface and a compensating flux in the opposite direction below [Smith, R. L., 1981]. A regression analysis indicates agreement of the magnitude of the fluctuating surface transport with that predicted from Ekman theory. The inferred magnitude of the lower layer transport is about a factor of two larger, indicating that the low frequency mass balance at that location is also three-dimensional.

The relation of the temperature fluctuations to the wind stress and to the alongshore velocity is illustrated nicely by the section EOF's. The first temperature mode (39% of the variance) has most of its structure near the surface and is well correlated with τ. Consistent with this, fluctuations in the temperature and the area of the cold water in the sea surface temperature maps are correlated with the wind stress [Stuart, 1981]. The area increases and the temperature decreases, as expected, for an increase in upwelling favorable wind stress. On the other hand, the second temperature EOF (22% of the variance) is well correlated with the first mode of v and reflects a fluctuating thermal wind balance. Most of the gradients in this T mode are concentrated around 200 m depth near the shelf break, indicating that the fluctuations in v tend to be relatively barotropic over the shelf and baroclinic over the slope [Brink et al., 1980].

This concentration of the temperature fluctuations associated with the fluctuations in the v field at a depth of about 200 m is consistent with results from Huyer's [1980] dynamic height and geostrophic velocity calculations. Time variations of the dynamic height extrapolated to the coast were found to agree with those of coastal sea level. The time dependence is associated primarily with variations in the isotherm depths between ~300-500 m. Huyer's [1980] computations are again extremely useful for the determiation of an offshore decay scale and the resulting estimate is 30-60 km. This scale agrees with estimates of the internal Rossby radius (discussed below).

The time-dependent balance in the depth-integrated alongshore momentum equation at mid-shelf (near mooring M) was investigated by Allen and Smith [1981]. Estimates for five terms were obtained; depth-integrated v_t and fu from current measurements (where subscript t denotes time derivative and f is the Coriolis parameter), pressure gradient p_y from alongshore differences of coastal sea level, bottom stress from near bottom velocity measurements together with a quadratic drag law, and wind stress. To assess the relative importance of these five terms, the time series were expanded in empirical orthogonal functions. It was concluded that the motion is dominated by an inviscid, unforced balance which, for the 0.1 to 0.2 cpd frequency band, is primarily between v_t and p_y.

While the alongshore current v at 15S is not correlated with the wind stress, it is highly correlated with v measured at other locations alongshore, i.e., at 10 and 12S. The time lags for maximum correlation indicate poleward propagation [Smith, R. L., 1978]. Cross-spectral analysis shows that this propagation occurs strongly in the 0.1-0.2 cpd frequency band and is nondispersive with an estimated speed in the range 200-300 km/day. The propagating fluctuations in v are associated with the first section EOF of the alongshore velocities. Consequently, the major share of the variance in the v field at 15S is associated with free coastal trapped waves. Based on observations of wave speed, offshore decay scale, the structure of alongshore velocity and temperature fluctuations, the v_t, p_y momentum balance, and on calculated theoretical mode shapes and wave speeds [Smith, R. L., 1978; Brink et al., 1978; Brink et al., 1980; Huyer, 1980; Allen and Smith, 1981; Brink, 1982a; Romea and Smith, 1983] these disturbances appear to have features much like a first mode free internal Kelvin wave with some modification in properties due to the bottom topography of the continental margin. An estimate of the internal Rossby radius at 15S obtained from the observed wave speed divided by the Coriolis parameter is about 55 km, which is greater than the shelf width and is consistent with the offshore scale found by Huyer [1980].

The source of these free waves is not known at present. Results from a forced wave model calculation by Brink [1982a], using observed winds and currents, suggest that the waves originate equatorward of 5S. It is theoretically possible that they are forced in the equatorial waveguide, propagate eastward along the equator and then southward along the South American coast. Romea and Smith [1983] have extended the data analysis concerning propagation to longer time and space scales using additional coastal sea level records. They find propagation between 2 and 17S and, with sea level data from the Galapagos Islands, "tenuous" evidence that the fluctuations come from the equatorial waveguide.

The question of what happens to free, first mode, coastal trapped internal Kelvin waves as they propagate poleward from low latitudes to mid-latitude regions where the internal Rossby radius is the order of, or smaller than, the shelf width, was investigated theoretically with an idealized two-layer model by Allen and Romea [1980]. It was found that, provided the wavelength was small compared with a distance (~1,100 km) dependent on model parameters (geometry, stratification, etc.), the cross-shelf modal structure of the wave would transform as it propagated poleward so that at mid-latitudes it would be close to a barotropic shelf wave. The numerical calculations by Brink [1982a], of coastal trapped wave mode characteristics as a function of latitude in a more realistic continuously stratified model, show essentially this same transformation for the first mode. For very low latitudes, on the other hand, e.g., near 5S, model results of Romea and Allen [1983] indicate that free or forced coastal trapped internal Kelvin waves may be characterized by vertical as well as poleward propagation.

Intermediate Frequency. Spectra of the current measurements at mid-shelf show peaks at diurnal and semi-diurnal periods, but no significant peak at the inertial frequency (0.52 cpd) [Brink et al., 1980]. Johnson [1981] found that, for the semi-diurnal motions, the alongshore velocity component had an amplitude larger by a factor of three than the cross-shelf component, typical kinetic energy levels of 6 erg/cm^3, and uniform phase over the shelf. The diurnal motions had the largest energy (~6 erg/cm^3) in the surface layer within 5 km of the coast, consistent with forcing by the diurnal sea breeze. Complex demodulation at the inertial frequency showed that the inertial motions were variable, but responded to wind forcing, with the largest amplitudes (~10 cm/s) near the surface over the inner shelf. Johnson [1981] concluded that, although spectral peaks at the inertial frequency were generally absent in the current meter records, inertial motions were intermittantly energetic.

Future Work

Although recent observations have led to greatly improved descriptions of shelf flow fields, a large amount of work remains in trying to understand the physics of several important processes. Some of the unresolved questions for each shelf region were brought out in the previous sections. Here we summarize briefly a few of the major problems that need future work and that are common to several shelf regions. In general, the work required will involve various combinations of the continued analysis of existing data, development of improved theoretical models, laboratory experiments, and further field experiments.

The dynamics of the coupling between the flow on the outer shelf and the offshore current and/or eddy regime over the slope certainly needs additional study. This is true not only in areas with strong offshore currents, such as the southeastern U.S. and west Florida shelves, with the Gulf Stream and Loop Current, respectively, but also in regions such as the Pacific Northwest, Gulf of Alaska, and eastern Bering Sea, with the relatively weaker California, Alaska, and Bering Slope Currents. This interaction is one part of the more general subject of the cross-shelf transport of mass, momentum, energy, salt, nutrients, etc. That remains one of the most important, yet most poorly understood, aspects of shelf circulation. Included are questions on the effects of density fronts and of topographic irregularities on cross-shelf transport, as well as questions on the differences between cross-shelf exchange processes in the surface frictional layer and in the water column below. Combined Eulerian-Lagrangian experiments should be helpful in investigations of these processes.

The alongshore variations of low frequency, seasonal mean, and annual mean currents on large alongshore scales (>100 km) are not well understood. For example, the nature of the coupling of the circulation on these time scales in the Middle Atlantic Bight, Georges Bank, Gulf of Maine, and Scotian Shelf system, needs additional research. In fact, the driving mechanisms for the annual mean currents on several shelves have not been unambiguously identified. The mean poleward current on the Peru shelf, which is opposite in direction to the persistent equatorward wind stress, provides one obvious example. Likewise, the dynamics involved in the mean southward currents in the Middle Atlantic Bight have not been fully resolved. It is noteworthy that, although the geographical locations differ considerably, the direction of these mean currents is the same as that of free, long, coastal trapped wave propagation. Although the significance of that fact is not known, it suggests that mechanisms with common roots may be involved. Related to general questions on maintenance of the mean alongshore flow are specific questions on the role in the alongshore momentum balance of the alongshore pressure gradient and the bottom stress. Those two fundamental forces play extremely important parts in even the simplest theoretical models. Establishing by measurements their contributions in actual shelf flow fields is an important task.

The dynamics of shelf currents driven by density differences due to river runoff and the interaction of these density driven flows with wind driven currents is another subject that needs additional work. Results from laboratory experiments will likely prove useful in these studies.

New information provided by the large number of recent observations has, in several instances, outrun our ability to rationalize it with theoretical models. The modelling aspect of shelf circulation will undoubtedly receive increased attention in the next few years. This should include the development of improved numerical models, with the incorporation of better representations of the surface frictional layer and of frictional and diffusive processes in general. In addition, the difficult, limited domain, open boundary condition problems inherent in constructing numerical models for shelf flow fields will have to be carefully dealt with. At the same time, there will be a need to further develop objective methods to test model predictions using available data sets.

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THE BENTHIC BOUNDARY LAYER AND SEDIMENT TRANSPORT

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1. Introduction

The major problems facing scientists studying marine boundary layer fluid mechanics in 1979 concerned the specific mechanismic operating in the regions of the top and the bottom of the boundary layer. At the top, the mechanisms of fluid escapement from the boundary layer and mixing into the ocean interior could only be postulated. Approximations of diffusivity across the top of the layer were being debated, the values of which were more articles of faith than physically validated estimates (cf. Armi, 1979; Garrett, 1979). At the bottom, the intermittent generation of stress (cf. Gordon, 1975; Heathershaw and Simpson, 1979) was held responsible for the wide variability observed in the drag coefficient, and the thickness of the viscous sub-layer was claimed to be greater than that traditionally found from laboratory studies (Caldwell and Chriss, 1979). On the bed, sediment transport models had to use Shields' values for entrainment, even though biological effects on sediment properties were recognized (cf. Rhoads, 1974). In the middle, the interaction of waves with currents and the effects of stratification by density and suspended sediment had been addresed only for special cases (cf. Smith, 1977).

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