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Variability of the coastal current and nutrient pathways in the eastern Gulf of Maine

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ABSTRACT

The eastern Maine coastal current flows southwestward, carrying cold and nutrient-rich waters along the coast from the tidally stirred eastern gulf toward the central and western gulf, where in summer the waters are warmer and more stratified. The current typically turns offshore before reaching Penobscot Bay, near the central coast, at a location determined largely by the distribution of dense slope water in Jordan Basin. The slope water, which enters the gulf as a deep inflow from the Atlantic Ocean, thus plays a major role in determining the intensity, direction and timing of the delivery of nutrients to the interior gulf. In this paper, we use data from two cruises in August 1987 to examine the variability and nutrient transport of the coastal current, especially to show the important physical linkages between the deep slope water, the structure of the coastal current, and its likely significant effect on biological productivity in the gulf.

1. Introduction

The extreme tides, cold waters, and high productivity for which the Gulf of Maine is well known are partly consequences of the shallow offshore banks that isolate and insulate the waters of the gulf from the Atlantic Ocean (Fig. 1). Generally, the gulf responds to external forcing as an inland sea connected to a deep ocean by a narrow channel, and not as a coastal ocean with an unrestricted seaward boundary, as the shape of the shoreline might suggest. The large tidal ranges in the gulf and especially its adjoining Bay of Fundy are a consequence of a near-resonance between the confined waters of the gulf and the semidiurnal Atlantic tidal wave, which sweeps past the mouth of the Northeast Channel with a period closely matching the period of natural response of the interior waters (Garrett, 1972). The near-resonance produces large horizontal tidal currents, especially over Georges Bank, the western shelf of Nova Scotia, and in the Bay of Fundy. In those areas, the amplified tidal currents keep the waters vertically mixed, resulting in low surface temperatures that are readily apparent in summer infrared images collected by satellites (Yentsch and Garfield, 1981). Bigelow (1927) first pointed out that the high productivity of gulf waters is

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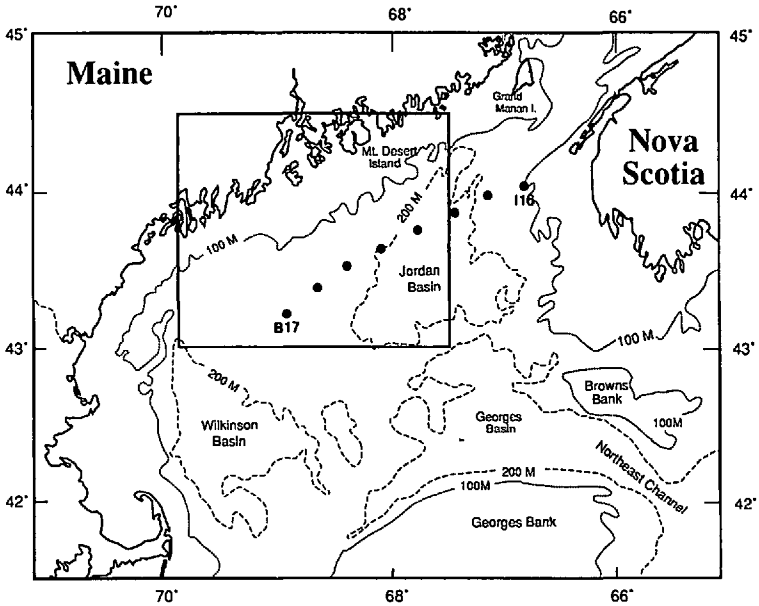


Figure 1. Map of the Gulf of Maine area, showing the principal basins of the gulf, the major offshore banks, and the deep channel that separates them. The 1987 survey area is shown by the inset box, for which a detailed map is given in Figure 2. Data from stations identified by the filled circles are shown in Figure 9; the station names are from a 1982 survey.

partly a result of the intense tidal stirring, which continually brings deep nutrients into the euphotic zone.

In summer satellite images, the tidally mixed water from the eastern gulf is evident in a plume of cold surface water extending southwestward along the coast from Grand Manan Island to the vicinity of Mt. Desert Island (Fig. 1). The cold water traces a nontidal coastal current that is associated with the contrast between buoyant nearshore water and deeper, more saline water confined in Jordan Basin (Brooks, 1985). A fraction of the coastal current turns offshore east of Penobscot Bay and enters a cyclonic gyre that develops in Jordan Basin in the summer. The remainder eddies over Jeffreys Bank and moves into the western gulf, where it contributes to the general counterclockwise circulation first described by Bigelow (1927).

The coastal current is important to productivity in the Gulf of Maine because it transports nutrient-enriched waters from the tidally mixed eastern gulf into the relatively warm and stratified surface waters of the western and central regions. Townsend *et al.* (1987) have calculated that about 44% of the nitrate entering the gulf as a deep flow through the Northeast Channel eventually is transported to the interior surface waters of the gulf by the coastal current. They also found that near-surface phytoplankton and zooplankton concentrations increased downstream in the current, and that the corresponding reduction in nitrate concentration was sufficient to support new primary production of 1.5 gm C m^{-2} d^{-1}, a significantly high value. Thus the high productivity of the gulf may depend upon three serially linked physical processes: (1) the injection or indraft into the gulf of nutrient-rich Atlantic slope water, which enters as an intermittent bottom flow through the Northeast Channel (Bigelow, 1927; Ramp *et al.*, 1985; Brooks, 1987); (2) the vigorous tidal mixing and associated upward vertical flux of nutrients into the surface waters of the eastern gulf, especially near the eastern Maine coast, where a deep extension of Jordan Basin allows the penetration of slope water into the region of intense tidal mixing (Bigelow, 1927; Loder and Greenberg, 1986); and (3) the coastal current transport of near-surface nutrients from the Grand Manan area into the warmer and stratified waters of the central and western gulf (Townsend *et al.*, 1987).

The distribution of deep slope water in Jordan Basin affects the structure of the coastal current, including the location at which it turns offshore, the fraction retained in the Jordan Basin gyre, and therefore the fraction available to transport cold water and nutrients westward into the rest of the gulf (Brooks, 1985). Thus the productivity stimulus for a large part of the gulf's surface waters may depend on the details of how Atlantic slope water spreads into Jordan Basin. With increasing clarity since Bigelow's (1927) first description of the process, it has become apparent that the slope water enters the Northeast Channel in bursts or events, perhaps related to the winds, but with a smoothed annual cycle that rapidly rises from a late winter minimum near zero to a maximum in early summer (Ramp *et al.*, 1985). The reasons for the annual signal are still uncertain, but it appears that the onset of summer stratification permits a relaxation of an offshore-directed deep pressure gradient that is sufficient to retard or even inhibit the inflow in winter (Hopkins and Garfield, 1979).

After entering the Northeast Channel, the slope water moves into eastern Jordan Basin and then spreads westward, generally following the topography of the inner edge of the basin. For the ranges of temperatures encountered, the water density is essentially controlled by its salinity (Brooks, 1985). Therefore the surface dynamic height reflects the deep slope water distribution, and part of the coastal current turns offshore around the western edge of the surface geopotential "low" associated with the advancing slope water. In the early stages of slope water inflow, such as the cases discussed here, the coastal current is directed offshore before it reaches the shoal topography of Jeffreys Bank, off Penobscot Bay (Fig. 2). When slope water floods Jordan Basin, typically later in the summer, an increased fraction of the coastal current is guided westward toward Jeffreys Bank.

The present paper shows the westward movement of the coastal separation point that occurred during a three week interval in the early summer of 1987, when the redirection of the coastal current was primarily controlled by the slope water distribution in Jordan Basin and not by the topography of Jeffreys Bank. We use hydrographic data and satellite imagery to suggest that the slope water steering mechanism may play an important role in determining the distribution of nutrients

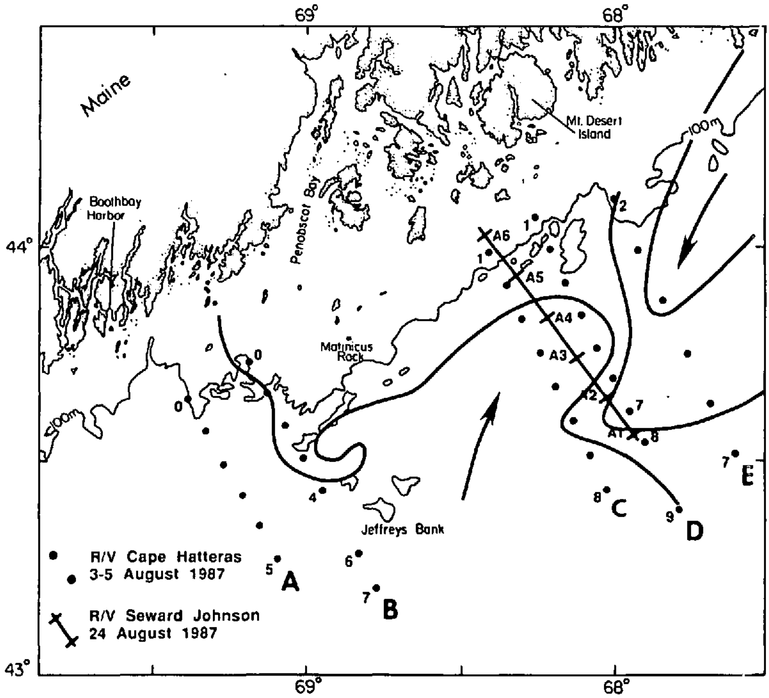


Figure 2. Base map showing the detailed survey area from the inset box in Figure 1. The station locations for the two cruises are shown. The solid contour curves represent gray-scale boundaries taken from a July 29, 1987 satellite infrared image, which reveals surface water temperature differences. The contours show cool surface water from the Maine coastal current moving offshore east of Mt. Desert Island and warmer water from the central gulf moving onshore south of Penobscot Bay. The arrows show the suggested direction of surface water movement. The 100 m depth contour (thin line) identifies Jeffreys Bank, off Penobscot Bay.

between the eastern and western basins of the gulf. Finally, we briefly examine older data from the gulf to show that the steering mechanism has been effective in other years, and we cite evidence indicating that a similar mechanism may be important in other coastal seas.

2. Evolution of the coastal current in 1987

We had two brief opportunities in 1987 to survey part of the Maine coastal waters; the first during August 3–5 in R/V *Cape Hatteras,* and the second about three weeks later during August 21–30 in R/V *Seward Johnson.* The cruise track and the station locations for the first cruise, shown in Figure 2, were chosen to survey the coastal current in the region of its offshore turn. An infrared satellite image from July 29 (Fig. 7a) indicated that the turning region was located east of Mt. Desert Island at the time, 70 to 80 km "upstream" of Jeffreys Bank. The second cruise was mostly concerned with biological sampling in the deeper waters of Jordan Basin (Townsend, 1987). However, a satellite image from August 16 revealed that, in the interim, the coastal current had spread westward toward the mouth of Penobscot Bay (Fig. 7b, c), so the cruise plan was rearranged to include a single hydrographic section extending inward toward the coast in the region between the "C" and "D" lines of the previous cruise, as shown in Figure 2. Thus the remotely sensed surface thermal structure of the coastal current can be compared with the measured hydrographic properties to illustrate the slope water steering mechanism.

For both cruises, conductivity, temperature and pressure (depth) data were obtained using a Neil Brown "Smart CTD," with an attached 12-bottle Niskin Rosette Sampler for water samples. The CTD data were recorded and averaged in 1 m depth intervals by the ship's data-logging computer, and standard software was used to compute salinities from the measured quantities. Chlorophyll concentrations were determined fluorometrically on the ship, using samples from the bottles. Nutient samples were also drawn from each bottle, then frozen for later analysis (primarily for dissolved nitrate concentration) at the Bigelow Laboratory.

The solid contour lines in Figure 2 show some of the gray-scale gradations that define the temperature patterns in the July 29 satellite image (Fig. 7a). At the time of the image, the local surface winds were weak and variable, with speeds ranging from calm to 6–8 knots (<15 m s^{-1}). The temperature contrast between the cold surface waters of the coastal current and the warmer waters of the central gulf can be clearly seen. The shape of the contours indicates southwestward and offshore movement of the coastal current and shoreward movement of warmer water; the latter appears to spread in both directions along the coast, with a suggestion of clockwise eddy recirculation off Mt. Desert Island. The apparent movement of the surface water is suggested by the arrows.

The gray-scale contour in Figure 2 that crosses the "D" section twice and ends at station D-9 also roughly defines the 15°C surface isotherm, independently contoured from the temperature measurements made 5 to 7 days later (Fig. 3). Both the satellite image and the ship measurements show that the coastal current turned offshore east of Mt. Desert Island during the first week of August. The warmer water moving toward the northeast inshore of the coastal current is clearly seen in Figure 3, contributing to the impression of a clockwise surface eddy south of Mt. Desert Island. During the first week of August, the surface winds were weak and variable except during a brief easterly weather disturbance on August 3, when 25-knot winds from the southeast may have encouraged shoreward movement of warm surface water; a comparison of the surface temperature patterns in Figures 2 and 3 suggests some strengthening of the clockwise eddy between the time of the satellite image (July 29) and August 5, when the easternmost section of the *Hatteras* survey was completed.

Contours of surface salinity and near-surface nitrate and total chlorophyll concen-

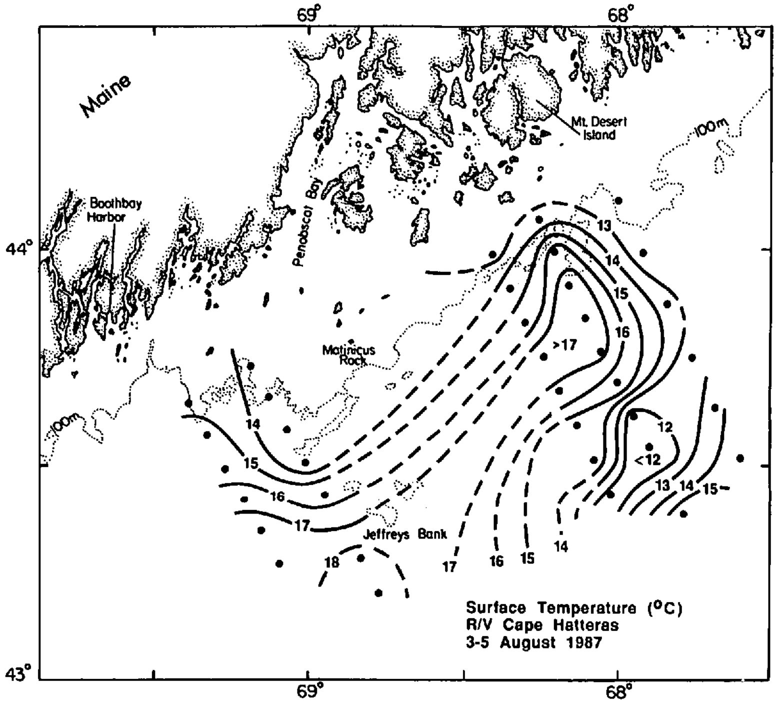


Figure 3. Contours of surface temperature (deg. C) for the *Cape Hatteras* cruise. The dashed lines indicate contouring uncertainty.

trations (unpublished cruise data) generally track the surface temperature contours of the cooler surface water in the redirected coastal current. Both the nitrate and the chlorophyll reached their largest values on the outer part of the "D" transect, with a sharp gradient or frontal zone near station D-6 separating the high values from much lower ones in relatively depleted inshore water.

Steering of the surface current by the deeper slope water is best illustrated by contoured depth of the 33 ppt salinity surface (Fig. 4). This salinity surface is chosen because it conveniently represents the distribution of slope water in the gulf (cf. Bigelow, 1927). In early August 1987, the slope water had not spread along the coast into the western part of Jordan Basin, but the western edge or "nose" of slope water is apparent on the eastern transect, where the 33 ppt surface rose to within 60 m of the surface (the heavy solid line traces the intersection of the 33 ppt surface with the bottom). The offshore turn of the cool surface water was tightly coupled with the deep slope water front, as a comparison of Figures 3 and 4 readily shows. The core of the surface current, defined by temperatures <13°C, was essentially bounded by the 70 and 90 m depth contours of the 33 ppt surface, showing that the offshore turn of the surface current was closely associated with the slope water distribution in the basin.

Since the slope water distribution effectively mirrors surface dynamic height, it is

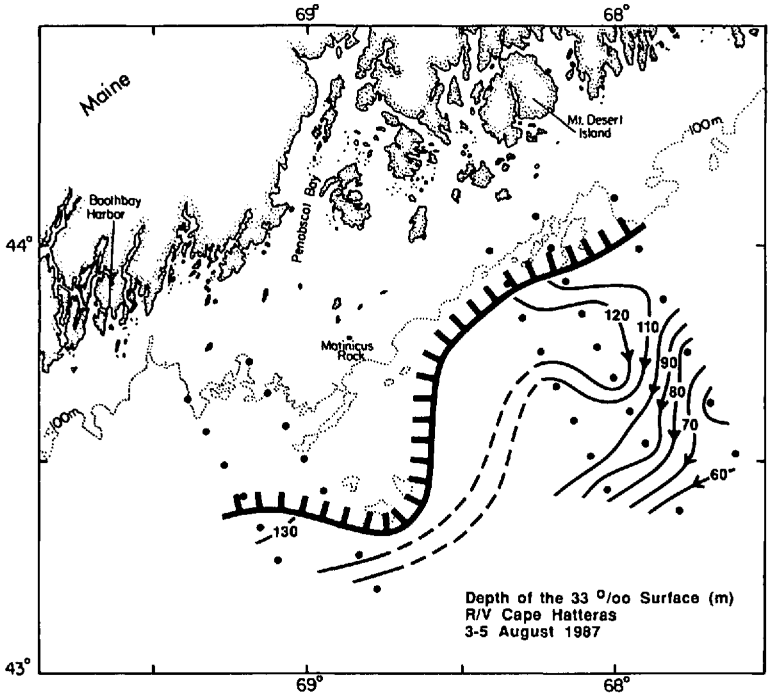


Figure 4. Contoured depth of the 33 ppt salinity surface (m) for the *Cape Hatteras* cruise, showing the distribution of deep slope water in the western part of Jordan Basin. The arrowheads show the indicated direction of upper level geostrophic flow, relative to deeper water, using salinity as a proxy for density. The coastal current turns southward across the two easternmost transects, with a clockwise eddy recirculation suggested inshore. The heavy curve with hachures shows the intersection of the 33 ppt surface with the bottom.

helpful to draw arrows on the deep salinity contours, showing the direction of geostrophic flow inferred for the near-surface layers relative to deep water (Fig. 4). This interpretation is equivalent to assuming a near-bottom reference depth, which has previously been shown to produce upper level geostrophic currents that are generally consistent with flow patterns inferred from tracer distributions, drifter tracks, and current measurements in the gulf (Brooks, 1985). In the present case, the upper level geostrophic flow pattern indicated by the deep salinity distribution is remarkably consistent with the surface flow inferred from the temperature patterns in Figure 3, and also with that suggested by the satellite image from July 29.

The contrasts between water types are more evident in vertical sections of temperature, salinity and geostrophic velocity from the easternmost transect, which sliced into the advancing slope water offshore of station E-4 (Fig. 5). The prominent mid-depth temperature minimum and relatively low salinity of Maine Intermediate Water (MIW) are evident on the inshore end of the section, but offshore of station E-4 the higher temperatures and salinities indicate the presence of slope water. Since the

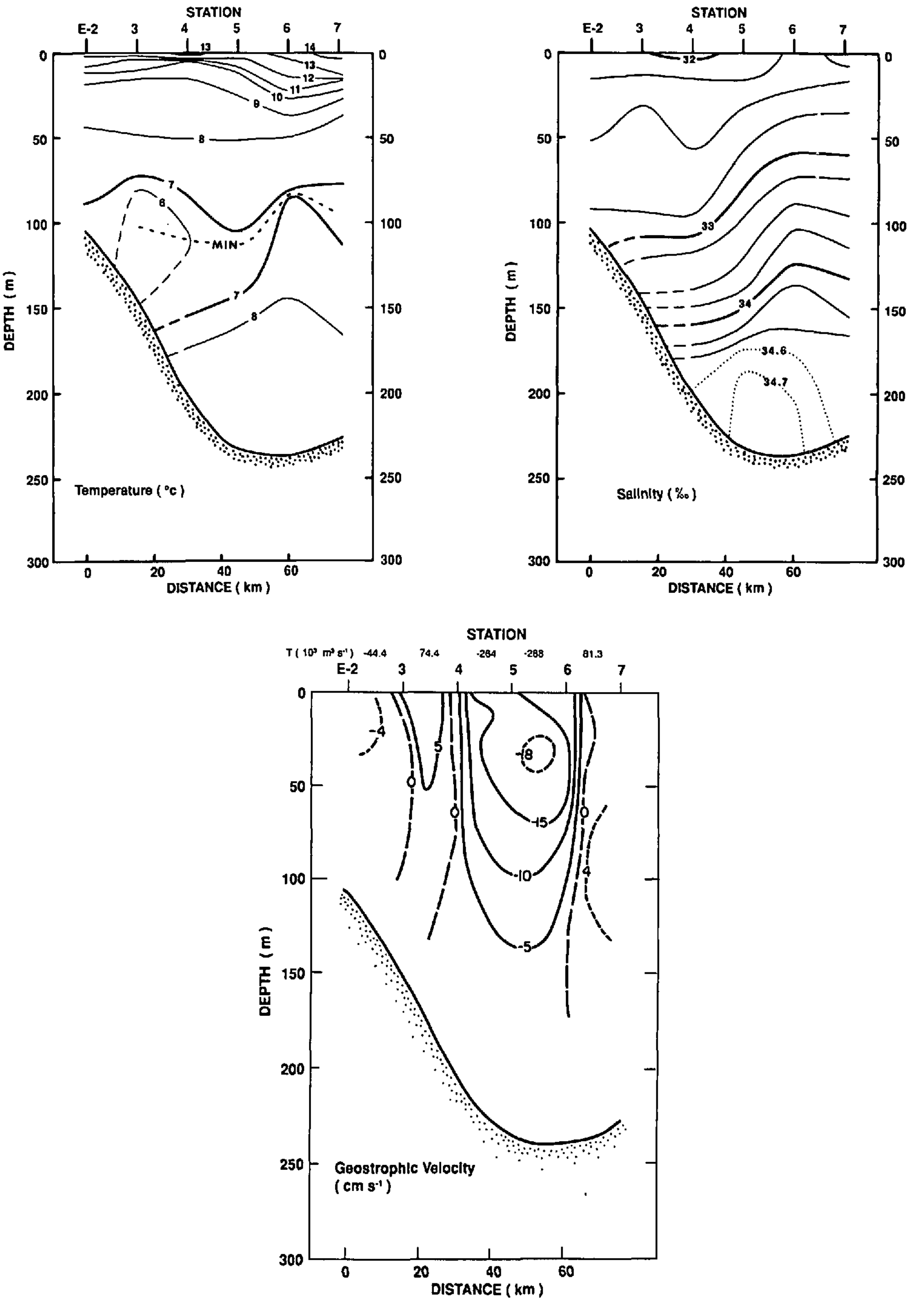


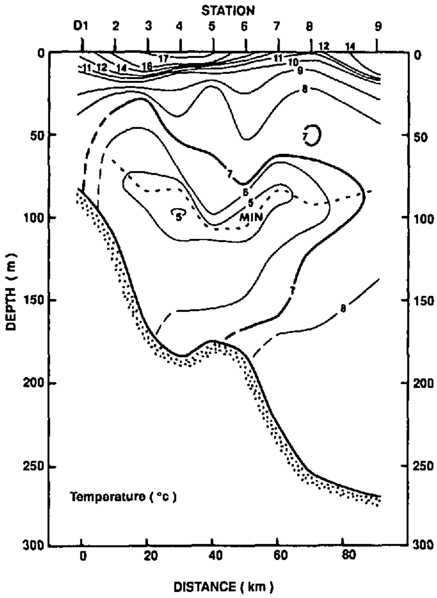
Figure 5. From the *Cape Hatteras* cruise on August 5, 1987, vertical sections of (a) temperature (deg c), (b) salinity (ppt), and (c) geostrophic velocity (cm s^{-1}) from the E-transect (cf. Fig. 2). The geostrophic velocity is computed relative to the greatest common depth between adjacent stations, which essentially gives the flow perpendicular to the section relative to the bottom. The zone of southwestward flow (negative contours) between stations 4 and 6 corresponds to the core of the coastal current as it crosses the section. The volume transport between stations is shown along the top axis of the figure.

isohalines are reasonable proxies for isopycnals here, the corresponding density section is not shown. The uplifted isohalines along the inner edge of the slope water correspond to a region of southwestward upper-level geostrophic flow between stations E-4 and E-6, defining the core of the coastal current (Fig. 5c). There are regions of weaker and less significant northeastward flow on each side of the southwestward core. The geostrophic velocities were calculated relative to the greatest common depth between adjacent stations, which essentially gives the flow relative to the bottom. This approach assumes that the slope water spreading takes place on a time scale long compared to the geostrophic adjustment time scale of the coastal current; i.e., that the geostrophic response of the coastal current is distinct from the processes responsible for slope water spreading. Ramp *et al.* (1985) found the winter slope water inflow fluctuations in the Northeast Channel to be most energetic in the 4–11 day period band, with longer periods indicated in the summer.

The net transport perpendicular to the "E" section was -440×10^3 m^3 s^{-1}, with about -550×10^3 m^3 s^{-1} carried between stations E-4 and E-6 in the core of the coastal current. The negative sign indicates southwestward flow. For comparison, based on two-week averages from current measurements in the Northeast Channel, Ramp *et al.* (1985) estimated the annual maximum inflow below 75 m to be about 400×10^3 m^3 s^{-1}, occurring in early summer. Our geostrophic transports are nearly synoptic estimates and thus are not directly comparable to Ramp *et al.'s* smoothed estimates, in which large inflow pulses of duration shorter than two weeks are masked; however, the coastal current transport appears to be important to the internal mass balance of the gulf.

We now shift our attention 20 km southwestward along the coast to the "D" transect, from which vertical sections of temperature, salinity and geostrophic velocity, and in addition nitrate and total chlorophyll concentrations, are shown in Figure 6. Here the MIW temperature minimum is evident across the entire section, indicating reduced influence of slope water, compared to the "E" section. Choosing the 7°C isotherm as a convenient bound, the MIW occupied depths from about 40 m to the bottom on the nearshore end of the section, thinning to the 80–100 m depth range at the offshore end. Offshore of station D-6, the near-bottom rises of temperature and salinity reflect the increasing influence of slope water as it penetrates from the east. The front-like zone of high salinity gradient, beginning at the bottom near station D-6 and then extending upward and offshore, distinguishes the deep slope water from the base of the fresher MIW.

The geostrophic velocity (Fig. 6c) shows a region of southwestward flow extending from station D-5 to beyond the offshore end of the section, with surface speeds >35 cm s^{-1} in a narrow, surface-intensified flow between stations D-6 and D-7. The coastal current crossed the section obliquely, so the geostrophic speeds underestimate the actual offshore speeds, which may have reached 50 cm s^{-1} near the surface. The total transport perpendicular to the transect offshore of station D-5 was -780×10^3 m^3 s^{-1}, almost twice that which crossed the E-transect. However, part of the



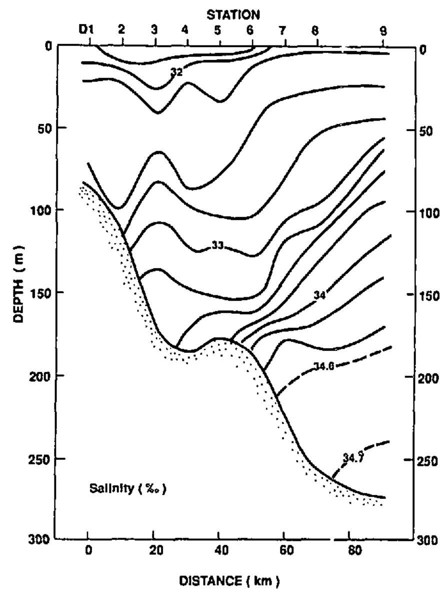


Figure 6. From the *Cape Hatteras* cruise on August 4, 1987, contoured vertical sections of: (a) temperature (deg. C), (b) salinity (ppt), (c) geostrophic velocity (cm s^{-1}), (d) nitrate concentration (μM), and (e) total chlorophyll concentration ($\mu g\ l^{-1}$) for the D-transect (cf. Fig. 2). The nitrate and chlorophyll values are from discrete bottle samples at nominal depths of 5, 15, 25, 45, 70, 100 m, and near-bottom. For clarity, some chlorophyll contours have been omitted near the surface at station D-7. The geostrophic velocity and transport estimates (the latter shown along the top axis) were computed relative to the bottom, with negative values representing southwestward flow across the section. The uplifted nitrate contours and enhanced near-surface chlorophyll values between stations 6 and 8 suggest upwelling and confinement in the eddy recirculation.

transport between stations D-6 and D-7 evidently came from a clockwise eddy recirculation of water that crossed the "D" transect toward the northeast between stations D-3 and D-5. The clockwise eddy is suggested in the satellite image (cf. Fig. 2), and it is also evident in the depth of the 33 ppt surface (Fig. 4), whose 110 m and 120 m contours indicate clockwise motion intersecting the "C" and "D" but not the "E" transects. Subtracting the eddy recirculation leaves a transport of about $-600 \times 10^3\ m^3\ s^{-1}$ in the coastal current offshore of station D-5, which is roughly comparable to the $-550 \times 10^3\ m^3\ s^{-1}$ estimated for the coastal current at transect "E." The weaker southwestward flow inshort of station D-3, with a subsurface core centered near 40 m depth, appears to be associated with a nearshore band of slightly fresher coastal water.

The nitrate and chlorophyll sections (Fig. 6) indicate upwelling and enhanced phytoplankton growth in the coastal current in the region where it turns offshore, especially in the zone of strong geostrophic flow between stations D-6 and D-8.

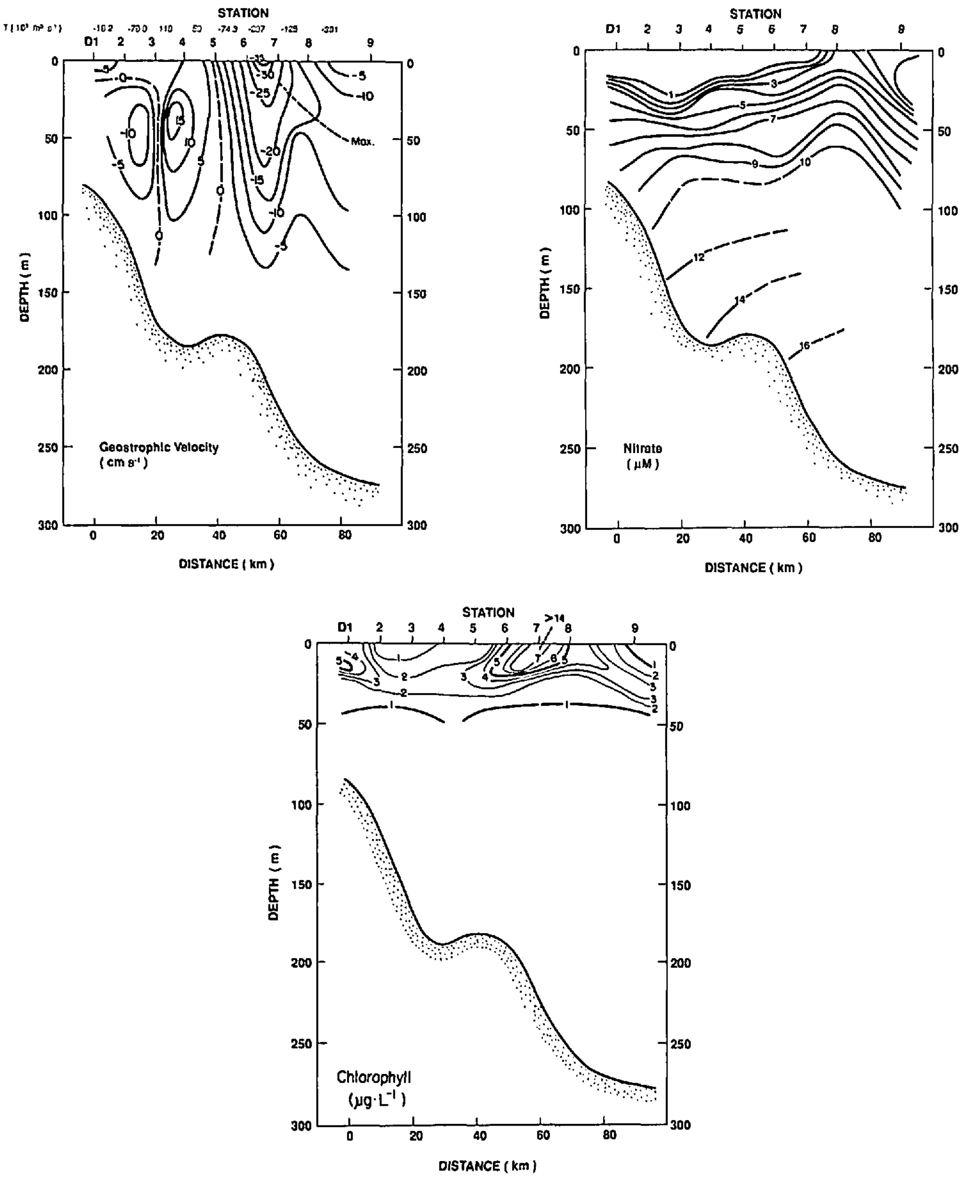


Figure 6. (Continued)

Upwelling, and not solely advection from the east, is indicated because the near-surface nitrate and chlorophyll values at station D-8 exceed those found at any of the other stations sampled during the cruise (unpublished data). In an earlier study, Townsend *et al.* (1987) found that nutrients were primarily advected from an eastern source, although in retrospect their results also show weaker evidence of upwelling in the coastal current where it turned offshore.

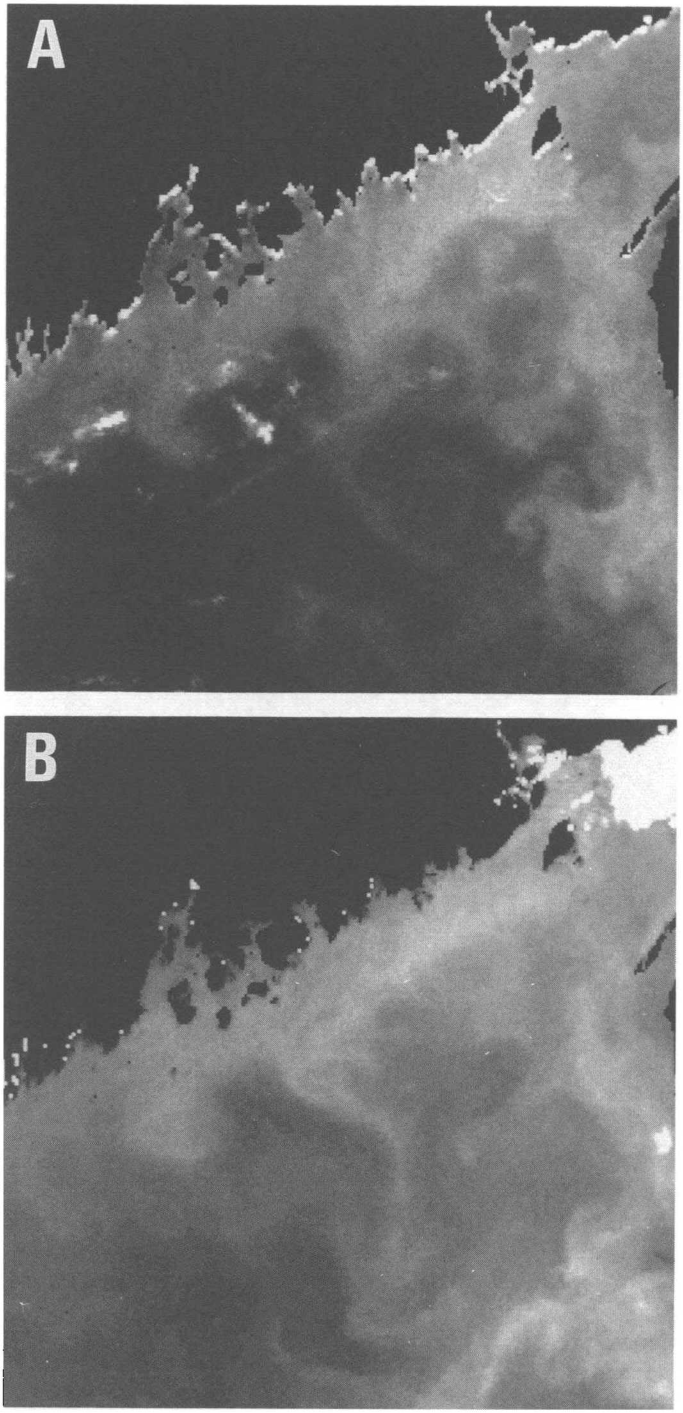


Figure 7. Satellite infrared images of the eastern Maine coastal region for (a) July 29, 1987, and (b) August 19, 1987. The lighter shades of gray represent cooler surface temperatures. Cold, tidally-mixed waters from the eastern gulf move southwestward in the coastal current, turning seaward and then back toward Jordan Basin before reaching Mt. Desert Island in the July image. In August, however, the coastal current extends farther westward before turning offshore and then back toward Jordan Basin. A stationary plume of cool coastal water extends offshore from Penobscot Bay in both images. Tidal mixing also keeps the surface waters cold over the Scotian Shelf, at the right in the images, and in the mouth of the Bay of Fundy, at the upper right. Part (c) shows the paths of the cool surface waters of the coastal current, determined from the satellite images, relative to the "E" transect of the *Cape Hatteras* cruise.

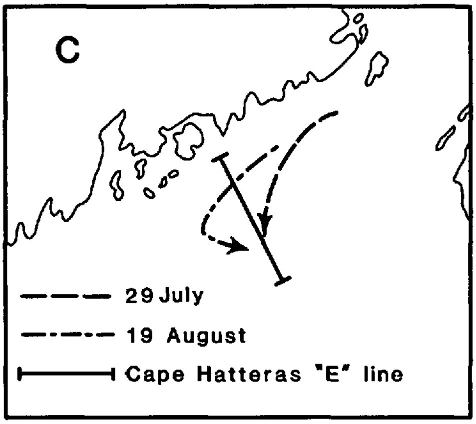


Figure 7. (Continued)

Uplifting of deeper nitrate-rich water near station D-8 is clearly indicated in the present case. At the D-transect, tidal mixing is not sufficient to homogenize the entire water column, as it is nearer the eastern end of the Maine coast. A more tenable explanation is localized upwelling driven by a nearshore divergence caused by the offshore-turning coastal current. Such a divergence could also induce a compensatory onshore movement of surface water from the central gulf, perhaps contributing to clockwise eddy motion between the redirected current and the coast. A coastal divergence would be encouraged by the upwelling-favorable southwest winds that often prevail in the summer months, but in July and early August 1987 the winds were variable, with easterly and northerly disturbances. The potential significance of the upwelling is indicated by the unusually high values of near-surface chlorophyll ($>14\ \mu g\ l^{-1}$ total pigments) found at station D-7 in the frontal zone between the eddy and the redirected coastal current. Eddy recirculation may have increased the residence time for phytoplankton growth in the upwelled, nutrient-rich waters.

3. An August sequel

A satellite image obtained on August 19 showed that the coastal current had spread westward along the eastern Maine coast during the three weeks intervening since the July 29 image (cf. Figs. 7a and 7b). To facilitate the comparison, in Figure 7c the locus of the cold core of surface waters in the coastal current is redrawn from each satellite image and shown relative to the "E" transect of the *Hatteras* cruise. In July, the coastal current turned seaward about 50 km east of Mt. Desert Island, whereas three weeks later, in August, it extended westward at least to Mt. Desert Island before turning offshore and then back toward the center of Jordan Basin. In the interval, the location at which the coastal current turned offshore moved westward at an average rate of a few km d^{-1}, not to be confused with the much greater indicated speed of surface water movement in the current.

The satellite images in Figure 7 also show a plume of cooler surface water extending offshore from the mouth of Penobscot Bay. Some of the plume water appears to be contiguous with the cooler waters from the east adjacent to the coast, indicating that an inshore branch of the coastal current may have escaped the offshore turn into Jordan Basin and continued west toward the bay. Such a branching of the coastal current is also suggested by the *Johnson* data, soon to be discussed. During the three week interval between the satellite images, the Penobscot plume remained stationary over Jeffreys Bank, allowing it to be readily distinguished from the offshore-turning coastal current, which was primarily associated with the developing hydrographic conditions in Jordan Basin.

The *Seward Johnson* coastal section, identified by station numbers A-1 through A-6 in Figure 2, began near station D-8 and ended near station C-1 of the earlier cruise. The temperature, salinity and geostrophic velocity from the *Seward Johnson* section are shown in Figure 8. Also shown are several superimposed isotherms and isohalines whose depths have been determined by linearly interpolating between the closest stations from the *Hatteras* "C" and "D" lines. A comparison shows the temperature and salinity increases inshore of station A-3 that resulted as the coastal current and slope water moved westward during the three week interval. For example, at station A-3, water with salinity >33 ppt was elevated about 50 m compared to its earlier state, while at the same station the MIW thickness (defined by the 7°C isotherm, as before) was reduced by about 25 m from above and 25 m from below. It also can be seen by comparing Figures 8 and 5 that the MIW temperature minimum increased by at least 1°C near station A-3, and the warming influence extended to within about 25 m of the surface between stations A-6 and A-3.

The geostrophic velocity perpendicular to the *Johnson* section (Fig. 8c) was southwestward between stations A-2 and A-3, with a maximum current speed of -18 cm s^{-1} at 20 m depth, and also between stations A-4 and A-6, with a maximum speed of -15 cm s^{-1} at the surface. The southwestward flow, especially near station A-3, is consistent with the location of the cold surface water shown in the later satellite image (Fig. 7b). The weakening of the southwestward flow between stations A-3 and A-4 may indicate a separation of the coastal current into two branches, one turning offshore and the other continuing along the coast toward Penobscot Bay, also suggested in the satellite images. The significance of the weak geostrophic flow reversal between the two branches is questionable, but some degree of branching is also suggested by the MIW distribution, whose upper bound (the 7°C isotherm) rises near station A-4 compared to adjacent stations. On the outer end of the section, a northeastward flow reaching 12 cm s^{-1} at the surface occurs where the August satellite image indicates movement of part of the redirected coastal current back toward the center of Jordan Basin. A station purposely located in this return flow showed surface chlorophyll levels in excess of 4 μg l^{-1}, which is unusually high for the normally oligotrophic offshore waters of Jordan Basin in the summer. The persistence of high

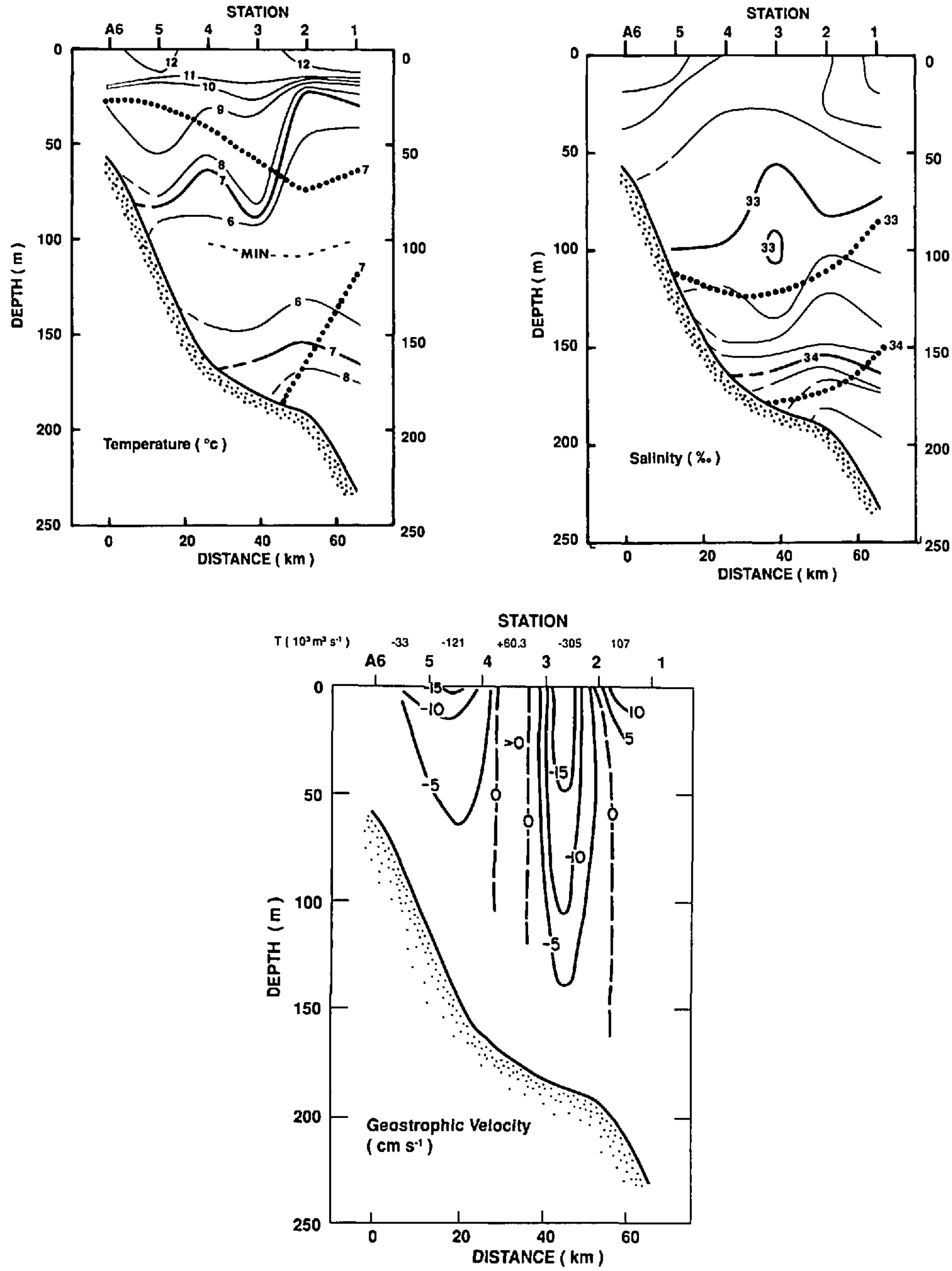


Figure 8. From the *Seward Johnson* cruise, contoured vertical sections of (a) temperature (deg. C), (b) salinity (ppt), and (c) geostrophic velocity (cm s^{-1}). The dotted curves overlaid for comparison are from the *Cape Hatteras* cruise, from which temperature and salinity values were linearly interpolated between stations from the C- and D-transects to compare with the closest station on the *Johnson* A-transect. At the time of the *Johnson* cruise, the core of the coastal current was between stations A-2 and A-3. The geostrophic velocity and transport estimates (the latter shown on the upper axis) are calculated relative to the bottom, with negative values indicating southwestward flow perpendicular to the section.

indicated productivity in the redirected coastal current shows its potential importance to the biological response of the offshore waters of the gulf.

The geostrophic transport across the *Seward Johnson* section was about $-400 \times 10^3\ m^3\ s^{-1}$ between stations A-6 and A-2. This value is comparable to but slightly diminished from the earlier transport estimates, which suggests a weakening of the coastal current in the three weeks between cruises, especially since the current apparently crossed the *Johnson* section more nearly perpendicularly than it did the earlier *Hatteras* section.

4. Other years and other seas

A re-examination of older data sets from the Gulf of Maine shows that the slope water steering mechanism just described has a degree of interannual persistence. For example, the eastern coastal region of the gulf was extensively surveyed in June, 1982. A fraction of the coastal current noted at that time moved southwestward over Jeffreys Bank, but the remainder turned offshore on a recurving path back toward the northeast, following the "low" in the surface dynamic topography that resulted mainly from high salinity slope water in Jordan Basin (Brooks, 1985). Figure 9 shows the salinity from a vertical section that runs generally parallel to the coast and crosses the deepest part of the basin. The section consists of stations taken from the cross-shelf transects of the 1982 survey; selected station locations are shown in Figure 1. Several isotherms are overlaid on the figure, but for brevity the corresponding temperature and density sections are not shown.

In the June 1982 section, the 33 ppt salinity contour rises to within 30 m of the surface at station G-14, near the center of Jordan Basin, and slopes down to a depth of 120 m at station C-15, closest to Jeffreys Bank. Over the 110 km separating the two stations, the isohaline slopes are consistent with a broad region of upper level offshore geostrophic flow relative to the bottom, with surface speeds of 10–20 cm s^{-1} between stations E-14 and G-14. The mid-depth temperature minimum of MIW is evident on the southwestern half of the section and the warming influence of the slope water can be seen in the basin. The two water types are separated by a sharp temperature front between stations E-14 and F-14, in the region of strongest offshore geostrophic flow. Clearly the slope water stood higher in the basin in June 1982 than in August 1987 (cf. Fig. 5), which suggests that the deep water inflow process began earlier or was more intense in 1982 than in 1987. Figure 9 shows that slope water with salinity >33 ppt extended west to Jeffreys Bank, but the inferred geostrophic flow indicates that offshore movement of the coastal waters was initiated well upstream of the bank, near the center of Jordan Basin, as also noted in 1987. Considered together with the cross-shelf transects shown by Brooks (1985), Figure 9 confirms that the circulation in the basin in the spring of 1982 is reasonably described as a cyclonic gyre whose center corresponds with the highest elevation of slope water in the basin, and whose inshore limb includes part of the coastal current referred to in the present paper. Compared

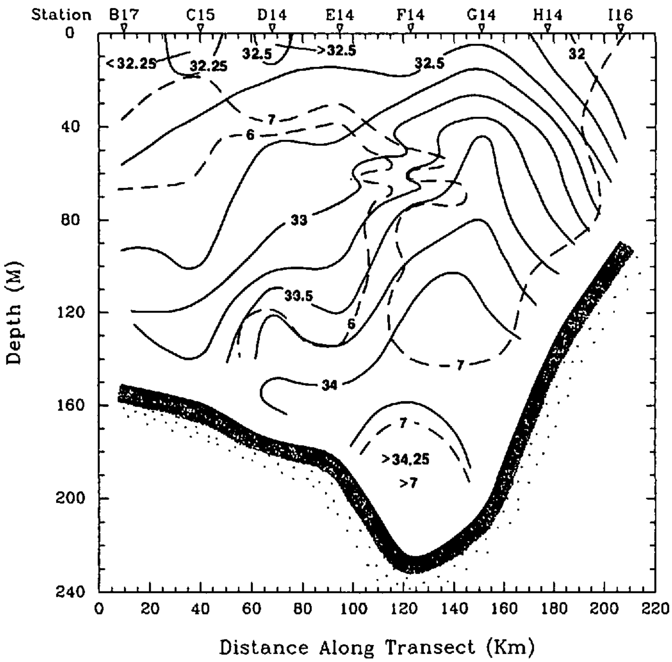


Figure 9. From 1982 data, an alongshore salinity section extending from a station southwest of Jeffreys Bank (B-17) to a station off the Scotian Shelf (I-16). The station locations are shown in Figure 1. The salinity contours (ppt) show an accumulation of slope water in Jordan Basin. The corresponding isopycnals (not shown) are similar to the isohalines, indicating an offshore geostrophic upper layer flow, relative to the bottom, between stations D-14 and G-14. The superimposed isotherms (deg. C) show the MIW, with its prominent temperature minimum, separated by a sharp front from the warmer slope water in Jordan Basin.

with 1987, the more advanced spread of the slope water in 1982 is consistent with the westward extension of the coastal current toward and over Jeffreys Bank noted for that year, but it is also clear that part of the coastal current in 1982 turned offshore east of Mt. Desert Island and contributed to the counterclockwise circulation in the basin, as it did in 1987 when the slope water intrusion was less extensive. It is interesting to note that Bigelow (1927), in his classic circulation schematic based on surface drifter data, included several streamlines showing a separation of the coastal current southeast of Mt. Desert Island, with offshore movement over Jordan Basin.

Finally, we note that steering of nearshore currents by denser bottom waters can be expected in many coastal or shelf seas in which there is a prominent contrast between oceanic waters and fresher shelf waters, so the mechanism described here is of more than local interest. In the Gulf of Maine, the contrast is particularly obvious and could be called estuarine. A similar but more complicated mechanism operates in the Georgia Bight, for example, where denser water from Gulf Stream meanders influences upper level water movements and nutrient distributions (Lee and Atkinson, 1983). In the western Gulf of Mexico, offshore jets of fresh and nutrient-rich coastal water sometimes result when the relatively dense waters in detached Loop Current rings reach the shelf edge (Brooks and Legeckis, 1982). An especially relevant parallel to the Gulf of Maine seems to occur in the Shelikof Strait, where the relatively fresh water of the Alaskan Coastal Current enters at one end and contrasts with more saline slope water that enters as a bottom current at the other end (Reed *et al.*, 1988). It appears that the deep slope water distribution influences the upper level currents and subsequent larval distributions in the Strait in ways similar to that described here for the eastern Gulf of Maine. A clearer understanding of the deep water steering mechanism in such areas would promote improved management of productive coastal seas in many parts of the world ocean.

5. Summary and conclusions

Slope water of Atlantic origin spread southwestward as bottom water in Jordan Basin during the first half of August 1987. Satellite and hydrographic data show that part of the eastern Maine coastal current turned offshore, following the deep contours of the slope water, as would be expected for a geostrophic upper level current. The location at which the surface current separated from the coast moved southwestward in concert with the expanding slope water in the basin. The separation occurred about 50 km to the northeast of Mt. Desert Island in the first week of August, whereas three weeks later it occurred south of the Island. The hydrographic surveys indicate that the westward extension of the current accompanied an extension or intrusion of deep slope water into Jordan Basin, supporting the hypothesis that the initial separation of the coastal current is geostrophically controlled by the spreading of deep slope water in the basin, and not by the shoaling topography of Jeffreys Bank farther west. Earlier data from the Gulf of Maine support this hypothesis. Upwelling induced by the divergence of the coastal current as it separates from the coast, perhaps aided by upwelling-favorable southwesterly summer winds, may be an important mechanism for bringing nutrients into the surface waters along the eastern Maine coast.

The biological impact of the redirected coastal current may be significant, because the current carries nutrients and phytoplankton from the tidally mixed eastern gulf into the central gulf surface waters, where weaker tidal mixing facilitates surface warming, stratification and hence higher production in spring and summer months. The location of the separation point may determine the division of the current's productivity stimulus between the eastern and western basins of the gulf. For example, an eastern or "early" separation during the stratified season would be expected to retard production in the west but enhance it in the east. The slope-water steering mechanism obviously depends on the annual cycle of slope water influx and spreading, which in turn is influenced by seasonal surface heating and cooling, river run-off, winds, and to some extent Gulf Stream rings, to name only the most apparent possibilities. A better understanding of the steering mechanism requires a numerical model capable of resolving the baroclinic structure of gulf waters while retaining realistic representations of complex bottom topography and boundary forcing. A similar mechanism may be important in other coastal seas that experience a prominent contrast between oceanic waters and fresher shelf waters.

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