



0278-4343(94)00054-9

Variability of water properties in late spring in the northern Great South Channel*

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Abstract—Regional CTD/ADCP surveys made in the northern Great South Channel (GSC) in late spring of 1988 and 1989 show different patterns of surface salinity in the extent of the freshwater plume east of Cape Cod. In April 1988, the surface plume was just beginning to form along the outer coast of Cape Cod, while 6 weeks later in the season in 1989, the minimum salinity was about 1.5 less, and a large pool of water fresher than 31.6 had pushed eastward over much of the northern GSC region. The difference in the amount of freshening between these two years is due primarily to the 6-week difference in the seasonal cycle and increased river discharge in 1989. The offshore spreading of the low-salinity plume was driven by the deeper circulation and upwelling-favorable winds.

The distribution of Maine Intermediate Water (MIW) also significantly differed between April 1988 and June 1989. In April 1988, the seasonal thermocline was just beginning to form, and the spatial structure of MIW was relatively uniform. In June 1989, a narrow core of temperature minimum water (with T_{min} in a range of 3.2–4.4°C) was found along the western flank of the northern GSC between 40 m and 120 m. This colder and fresher water spread to mix with the interior MIW as the core flowed southward into the central GSC.

Hydrographic data plus satellite sea-surface temperature images showed a relatively permanent continuous thermal front (with a 10-km cross-isobath variation) along the eastern flank of Nantucket Shoals, across the northern shallow region of the GSC and along the northwestern flank of Georges Bank, which separated the well-mixed water over the shallow region of the GSC from stratified water in the center of the northern GSC. Comparison of the location of this front with theoretical predictions by LODER and GREENBERG [(1986) *Continental Shelf Research*, 6, 397–414] suggests that enhanced tidal mixing due to the spring–neap cycle is important in determining the relative balance between buoyancy import and tidal mixing in the GSC region.

1. INTRODUCTION

THE Great South Channel (GSC), located in the southwestern Gulf of Maine (GOM), runs approximately north–south, with a sill depth of about 50 m near 40°45' N to separate the mid- and outer continental shelf to the south from the deeper GOM to the north, where depths exceed 150 m. To the east of the GSC is Georges Bank, a large elliptical submarine bank with a minimum water depth of less than 40 m on the top of the bank. To the west is Nantucket Shoals, a large shallow region extending southeast from Nantucket Island.

Water properties in late spring in the western GOM are mainly characterized by: (1) a

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nearshore band of low-salinity water (LIMEBURNER and BEARDSLEY, 1982), (2) deep Maine Intermediate Water (MIW, BROWN and BEARDSLEY, 1978; HOPKINS and GARFIELD, 1979); and (3) Maine Bottom Water (MBW, MOUNTAIN and JESSEN, 1987; RAMP *et al.*, 1985). Based on two relatively complete regional hydrographic surveys conducted in the western GOM during 11–12 May 1976 and 24 May–7 June 1979, LIMEBURNER and BEARDSLEY (1982) reported a surface low-salinity plume spreading along the western coast of the GOM and into the northern GSC region. This plume is believed to originate from spring river discharge into the GOM, primarily along the Maine and New Brunswick coasts and, to a lesser extent, the freshwater inflow from the Scotian shelf (SMITH, 1983). The MIW is formed by surface cooling during winter and surface heating during spring. The winter overturning due to surface cooling and wind mixing leads to vertical mixing between the cold, fresh surface water and warm, saline deep water within the GOM. During the following spring, surface warming re-stratifies the upper surface layer to cause the relatively prominent mid-depth temperature minimum water mass called MIW. MBW is formed by mixing between MIW and the influx of warm, high salinity Slope Water through the Northeast Channel east of Georges Bank.

In spite of our knowledge of water properties in the western GOM, there are still many unanswered questions about water property variability and the underlying subtidal circulation in this region. For example, does the surface low-salinity water in spring flow southward along the east flank of Nantucket Shoals through the GSC or does it bifurcate as suggested by LIMEBURNER and BEARDSLEY (1982) and also flow northeastward along the northern flank of Georges Bank? Noting that the T - S envelope of MIW was similar both north and south of the 50-m deep sill of the GSC, HOPKINS and GARFIELD (1979) suggested a southward flow of MIW through the GSC to the New England shelf as a conservative water mass. However, since strong tidal mixing is usually observed over the GSC sill, how can MIW flow through the GSC without significant mixing with surface water?

As part of the South Channel Ocean Productivity Experiment (SCOPEX), two regional CTD/ADCP surveys coupled with satellite-tracked drifter deployments were conducted in the northern GSC during 26–29 April 1988 and 6–12 June 1989 [Fig. 1(a) and (b)]. The principal objectives of these measurements were (a) to characterize regional water property variability, (b) to investigate the structure of the low-salinity surface plume observed east of Cape Cod in late spring, (c) to locate the front or boundary caused by tidal mixing between the vertically well-mixed water over Nantucket Shoal, the GSC, and Georges Bank and the stratified water in the deeper northern GSC, and (d) to determine the three-dimensional structure of the subtidal flow field in the northern GSC and its relationship to the high zooplankton concentrations found there in spring. A reduced regional CTD/ADCP survey was also made during 18–21 May 1989 on the western flank of the GSC east of Cape Cod to see if the surface low-salinity plume was present then and thus help determine the seasonal evolution of the plume during spring [Fig. 1(c)]. To help interpret these measurements, we have also obtained (a) the CTD/ADCP data collected on the May and June 1989 SCOPEX small-scale biological surveys, (b) satellite sea-surface temperature (SST) images for spring 1989, (c) hourly wind data collected on local NOAA environmental buoys (Fig. 1), (d) the monthly freshwater discharge data for key Maine, New Hampshire, and Massachusetts rivers, and (e) the monthly surface salinity distribution data collected along the Boston–Halifax transect through the NOAA Ship Of Opportunity Program [SOOP, see Fig. 1(d)].

In this paper, we will present a detailed description of the water property distributions

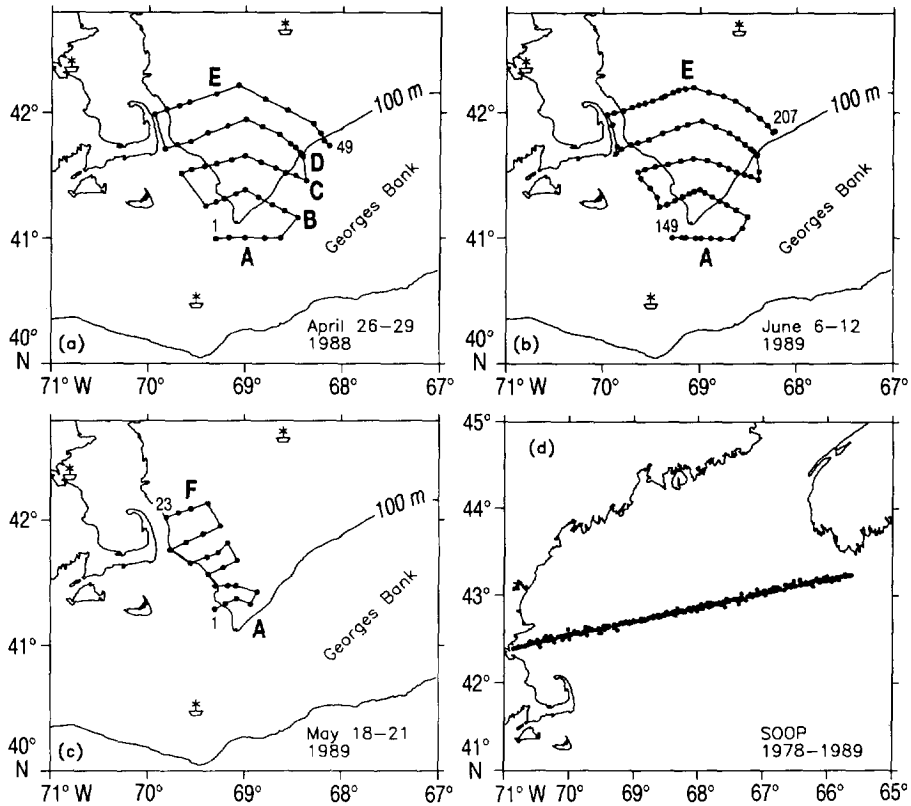


Fig. 1. Locations of regional SCOPEX CTD survey stations taken during (a) 26–29 April 1988, (b) 6–12 June 1989, and (c) 18–21 May 1989. ● is an individual CTD station and integers in the maps are the CTD station numbers. The cross-channel transects running from the western flank of the northern GSC to its eastern flank are labelled A–E from south to north for the April 1988 and June 1989 surveys and A–F for the May 1989 survey. (d) Locations of surface salinity stations taken on the Boston–Halifax NOAA Ship Of Opportunity Program transects during 1978–1989. ☼ indicates the position of NOAA environmental buoys.

observed in the northern GSC in late spring 1988 and 1989. In particular, we will focus on the structure and evolution of the surface low-salinity plume, MIW, and the position of the tidal mixing front. Based on the results of the SCOPEX drifters and ADCP current measurements [which are described in detail separately by CHEN *et al.* (1995)], we will present here a simple conceptual model of how wind and interfacial friction can form and move the surface low-salinity plume east of Cape Cod in late spring. The influence of the observed water property structure and subtidal circulation on zooplankton distribution and patchiness is described separately by WISHNER *et al.* (1995).

2. SURFACE LOW-SALINITY PLUME EAST OF CAPE COD IN LATE SPRING

Mountain and Manning (personal communication) studied the annual and interannual variability of water properties in the GOM using the NOAA MARMAP 1977–1987 10-year hydrographic data set. They found that the variability of the low-salinity surface water

in the western GOM is due primarily to the seasonal variability of local river discharge, with a maximum in discharge occurring in late spring and a minimum in winter. Hydrographic data taken in the SCOPEX 1988 and 1989 CTD surveys provide several snapshots of the distribution of surface salinity in the northern GSC in late spring. The variability of the low-salinity surface plume is described in the following within the context of the conceptual model of a seasonal springtime evolution of the salinity field in the western GOM described by Mountain and Manning, with the understanding that the SCOPEX hydrographic data considered in isolation cannot be used to prove if such a seasonal cycle and springtime variation exist.

The surface salinity patterns observed in late April 1988 and May–June 1989 differ markedly in the extent of the low-salinity plume which occurs over the western flank of the northern GSC in spring (Fig. 2). In April 1988, the low-salinity plume was just beginning to form and was confined to the east coast of Cape Cod, while in May–June 1989, three to six weeks later in the year, the plume was spreading offshore from Cape Cod and finally covered most of the northern GSC as a large pool from Wilkinson Basin to the southern sill of the GSC between Nantucket Shoals and Georges Bank. The minimum salinity of the plume was about 32.0 in April 1988 while it was 0.4–1.2 fresher in May–June 1989. Such a big difference is also evident in the distribution of monthly surface salinity across the western Gulf of Maine obtained on the Boston–Halifax SOOP transects during 1988 and 1989 (Fig. 3). Taking the 32.0 contour as a reference boundary of fresher water, we found in 1988 that the fresher water was beginning to appear in Massachusetts Bay in early April and extended to Wilkinson Basin after June, while in 1989, the fresher water was beginning to enter Wilkinson Basin in late April although it entered Massachusetts Bay in mid-April. The volume of low-salinity water found in Massachusetts Bay from May to August was considerably larger in 1989 than in 1988.

The difference in the amount of freshening observed in the northern GSC between April 1988 and June 1989 is due to the increased local river discharge in 1989 and a 3–6 week time difference in the seasonal cycle between the two surveys. Distributions of monthly surface salinity along the SOOP transect during the last 10 years from 1978 to 1989 clearly show two regions of fresher water, one in Massachusetts Bay and Wilkinson Basin and the other on the Scotian Shelf (e.g. Fig. 3). Larger river discharge was found from New Brunswick to Cape Cod in 1989 than in 1988. For example, the maximum Penobscot River monthly averaged runoff was about $700 \text{ m}^3 \text{ s}^{-1}$ in April 1988 while it increased to $1100 \text{ m}^3 \text{ s}^{-1}$ in May 1989 (see Fig. 4). Information on the structure of the low-salinity surface plume is also available from previous hydrographic surveys taken in May 1976 (LIMEBURNER and BEARDSLEY, 1982) and in July 1987 (GARRISON and BROWN, 1989). Taking the 32.0 contour as a reference boundary of this low-salinity plume and ignoring its inter-annual variability, we find that the plume extends eastward with time in a seasonal cycle from just east of Cape Cod to the western flank of Georges Bank (see Fig. 5). Although interannual variability in river runoff and other factors may change the rate of spreading, these data taken together support our simple conceptual picture of a freshwater plume which spreads eastward in the northern GSC and becomes fresher as time advances from April to June.

3. VERTICAL STRUCTURE OF WATER PROPERTIES IN THE NORTHERN GSC

The vertical structure of water properties was described using the CTD measurements along the northernmost transect in the April 1988 and May–June 1989 surveys (Figs 1 and

6–8). In April 1988, the seasonal thermocline was just beginning to form, and the freshwater plume can be clearly identified in the vertical salinity distribution over the western flank of the northern GSC, where a strong salinity front was located. The maximum depth of the front was about 40 m, and its cross-shelf scale was about 40 km extending east from Cape Cod. Farther offshore was located Maine Surface Water (MSW, see SMITH, 1983), which was characterized by a range of temperature from 5.6°C at the surface to 4.8°C at 40 m, and a relatively uniform salinity between 32.6 and 32.7. In May 1989, a strong thermocline and halocline were observed in the upper 40 m, while the thermocline and halocline were more concentrated in the upper 30 m in June 1989, 3 weeks later in the seasonal cycle. At that time, the low-salinity plume had a surface salinity less than 32.0 and extended far offshore east of Cape Cod into the central GSC region, although the freshest water was still found in a narrow surface band next to Cape Cod with a cross-shelf scale of 40 km and a vertical scale of 40 m. Unlike April 1988 and May 1989, the low-salinity plume exhibited little vertical temperature and salinity structure in the upper 20 m in June 1989, especially in the interior region away from the coastal tidal mixing front. The lack of much vertical stratification there may be a result of surface wind mixing.*

In all three surveys, Maine Intermediate Water (MIW) occupied the middle of the water column beneath the thermocline and halocline. In both 1988 and 1989, MIW was characterized by a temperature minimum less than 4.8°C centered between 40 m and 150 m and an intermediate salinity (see Figs 6 and 7). Tracing the core of temperature minimum water by defining the 4.8°C contour as the boundary of the MIW for the 1988 and 1989 surveys, we found that the vertical thickness and horizontal extent of MIW tended to decrease eastward and southward, respectively, and finally disappeared over the sill of the GSC where the water was relatively well mixed in the vertical. This implies mixing between the MSW and MIW as the water flowed southward along the western flank of the northern GSC. Moreover, in April 1988, the spatial structure of MIW was relatively uniform, while in May 1989, a narrow core of temperature minimum water with temperature just less than 4.0°C and salinity of about 32.9 was found along the western flank of the northern GSC between 40 and 150 m. In June 1989, 3 weeks later, this temperature minimum core had become much cooler (in a temperature range of 3.2–4.0°C) and fresher (in a salinity range of 32.4–32.6). This colder, fresher, and lighter water seemed to be isolated from the interior temperature minimum and spread offshore to mix with the interior MIW as the core flowed southward in the northern GSC (see Fig. 7 and CHEN, 1992). Although the upstream source of this temperature-minimum water is unclear, we speculate that this very cold and fresh water mass is formed in late winter along the western GOM coast and flows southward into the northern GSC as an intermediate-level buoyancy-driven current.

Whether or not MIW can pass southward through the GSC onto the New England shelf was examined based on T - S diagrams at stations made along the southernmost transects (labelled A in Fig. 1) in the April 1988 and June 1989 surveys. In April 1988, the water was weakly stratified between Stas 1 and 3 on the western side of transect A (see Fig. 1). Nearly straight T - S curves were observed at Stas 1–3, respectively, with a temperature and

*Note: the vertical gradient found in salinity but not in temperature in the upper 20 m in Fig. 6 is due to the smaller contour plotting resolution used for salinity. Typical vertical salinity and temperature differences in the top 20 m were 0.01–0.02 and 0.02–0.04°C, respectively.

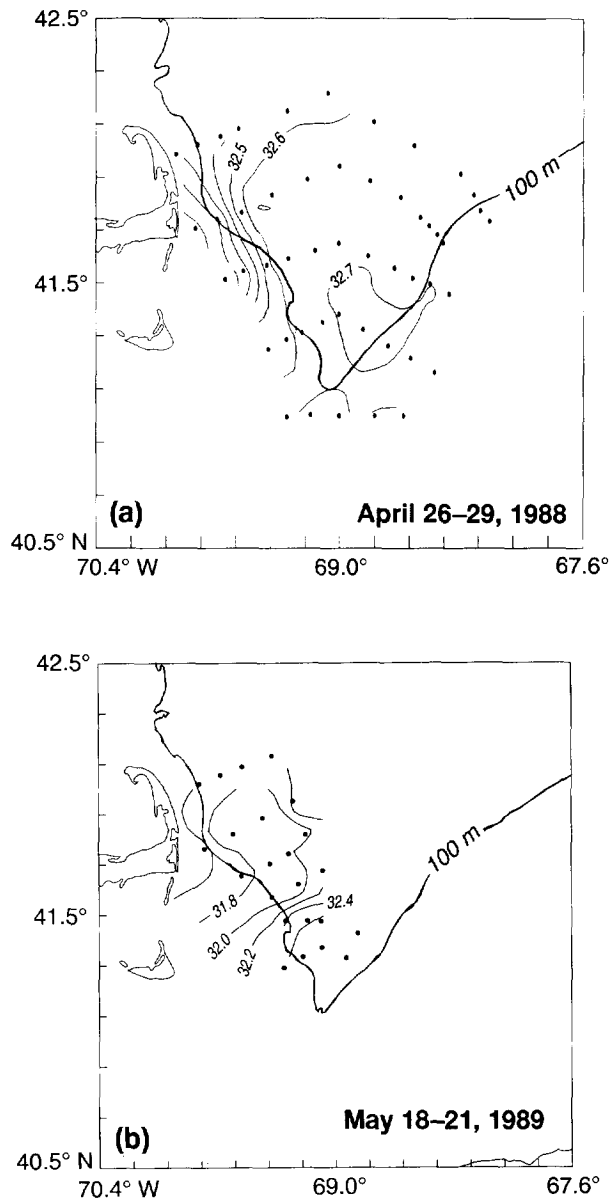


Fig. 2. (a) Surface (2 m) salinity distributions during 26-29 April 1988. The heavy solid line is the 100-m isobath. (b) Surface (2 m) salinity distributions during 18-21 May 1989. The heavy solid line is the 100-m isobath. (c) Surface (2 m) salinity distributions during 6-12 June 1989. The heavy solid line is the 100-m isobath.

salinity of 5.2-6.1‰ and 32.5-32.6 at the surface and 5.1-5.3‰ and 32.7-32.8 at the bottom [Fig. 8(a)]. These nearly linear $T-S$ curves can be traced northward to transect B only in the upper 50 m where MSW was dominant, implying that only a small part of MIW shallower than 50 m may pass through the GSC as water mixed with MSW in April 1988. In

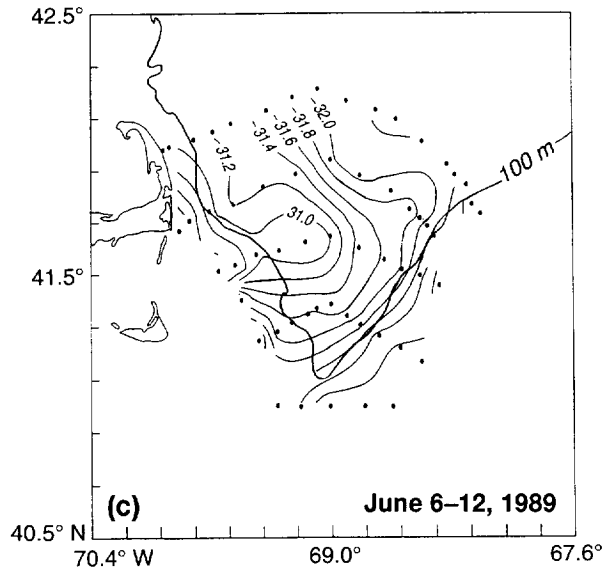


Fig. 2. (Continued)

June 1989, a big difference in the T - S curve occurred between Stas 150 and 151 on transect A, indicating two significantly different mixed water masses: (1) the colder and fresher mixed water of MIW and MSW to the west; and (2) the relatively warm and saline Georges Bank water to the east [Fig. 8(b)]. The existence of these two water masses was consistent with the low-pass ADCP currents and trajectories of satellite-tracked drifters where southward and northeastward currents were found on the western and eastern sides of transect A, respectively (see CHEN *et al.*, 1995). In contrast with HOPKINS and GARFIELD's (1979) suggestion, mixing plays an important role over the sill of the GSC so that no original MIW can pass south through the GSC without mixing with MSW during late spring.

The deepest water found in these surveys was Maine Bottom Water (MBW), caused by a westward penetration of warm and saline Slope Water from the Northeast Channel (MOUNTAIN and JESSEN, 1987). The MBW is characterized here as that water lying beneath MIW between about 120 m and the bottom in which both temperature and salinity increase with depth (Figs 6 and 7). In addition, the MBW seemed to be more stratified in April 1988 in the north central GSC than in June 1989.

4. TIDAL MIXING FRONT IN LATE SPRING IN THE NORTHERN GSC

When wind mixing, horizontal advection, and fresh water input are ignored, local vertical mixing in coastal waters is controlled primarily by a competition between the surface buoyancy flux due to solar heating and kinetic energy dissipation caused by oscillating tidal currents over the bottom. When tidal energy dissipation is stronger than the buoyancy input, the water will be vertically well mixed. In turn, the water will remain stratified when the surface buoyancy flux is dominant. The transition zone between well-mixed and stratified regions (called the tidal mixing or tidal front) should be located where these two processes are balanced. Based on this argument, SIMPSON and HUNTER (1974)

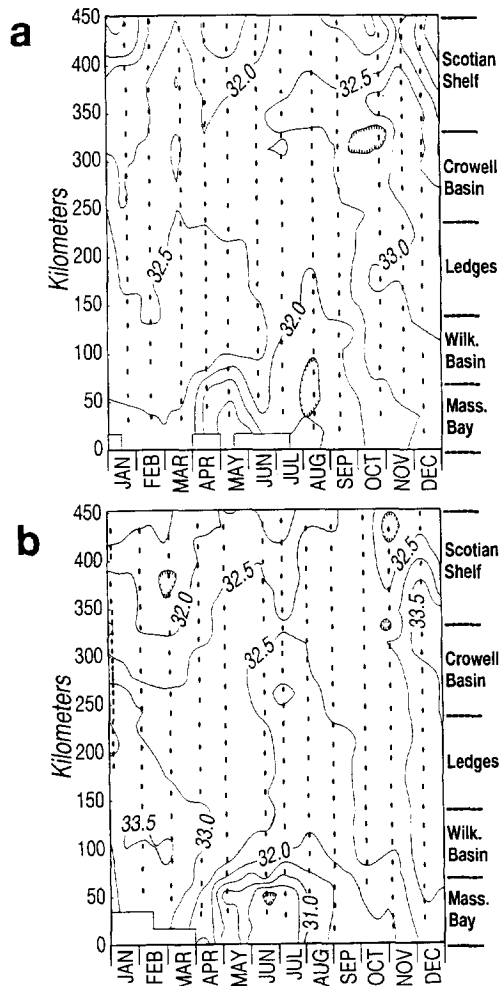


Fig. 3. Surface salinity distributions along the Boston-Halifax SOOP transects during (a) 1988 and (b) 1989. These maps were provided by J. W. Jossi of the NMFS Northeast Fisheries Center, Narragansett Laboratory.

suggested that the existence and position of the tidal mixing front can be predicted using a particular value of the ratio of potential energy required for complete vertical mixing and the rate of dissipation of mechanical energy due to tidal currents (so called mixing efficiency). In an area of constant surface heat flux, thermal expansion coefficient, specific heat and bottom friction coefficient, this ratio is proportional to h/U^3 or h/D , where h is the local water depth, U a characteristic tidal velocity, and D the rate of tidal energy dissipation.

GARRETT *et al.* (1978) applied this criterion to the Gulf of Maine using available hydrographic data and vertically averaged tidal currents from the numerical tidal model developed by GREENBERG (1979). They found that regions of $\Delta\sigma_t \leq 0.5$ in the upper 40 m over Georges Bank and off southwestern Nova Scotia correspond reasonably well to regions of $\log_{10}(h/D) < 2.0$, and predicted that a definite transition between stratified and

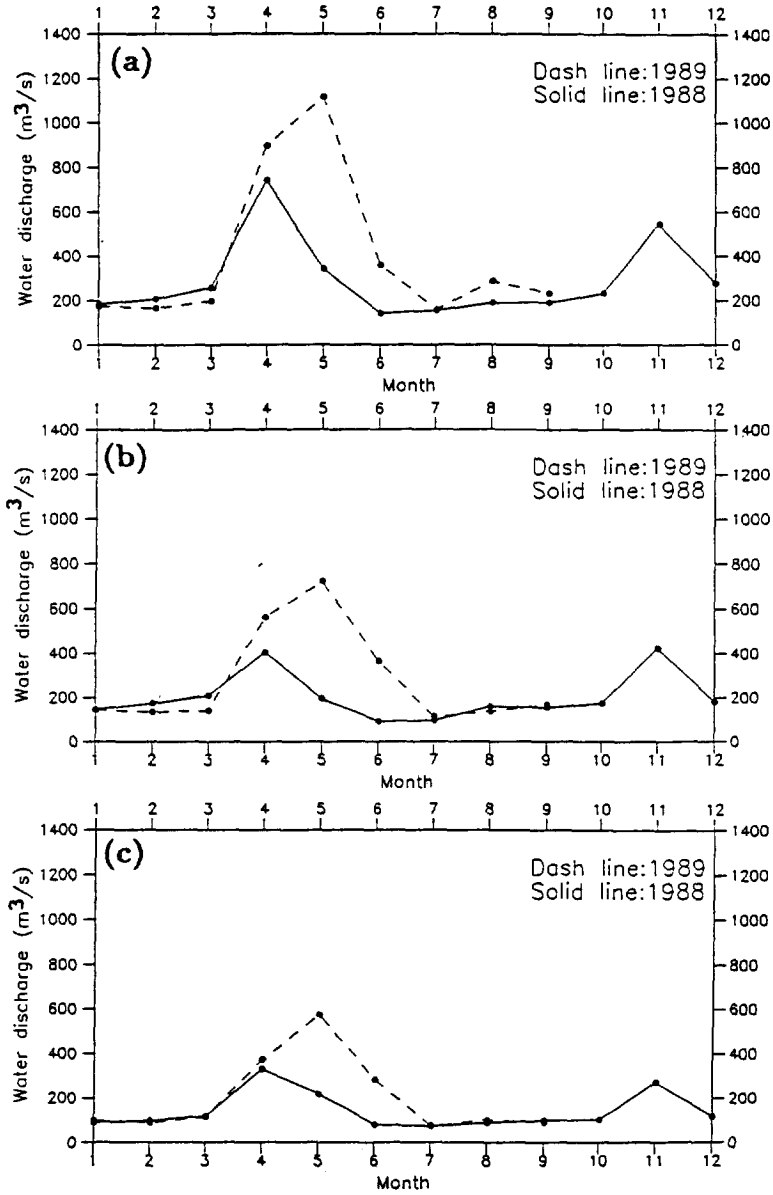


Fig. 4. The 1988 and 1989 monthly averaged freshwater discharge from (a) the Penobscot River at Eddington, (b) the Kennebec River at North Sidney, and (c) the Androscoggin River near Auburn.

well-mixed regions occurred at $\log_{10}h/D = 1.9$. LODER and GREENBERG (1986) examined the effects of wind mixing, oscillatory advection in a tidal cycle, and monthly and fortnightly variation in the strength of tidal mixing on the prediction of tidal front position. Adding the wind stress due to a 10-m wind speed of $1-5 \text{ m}^{-1} \text{ s}$ (a climatological mean wind speed in June and August in the GOM, SAUNDERS, 1977) into the energetic argument, they

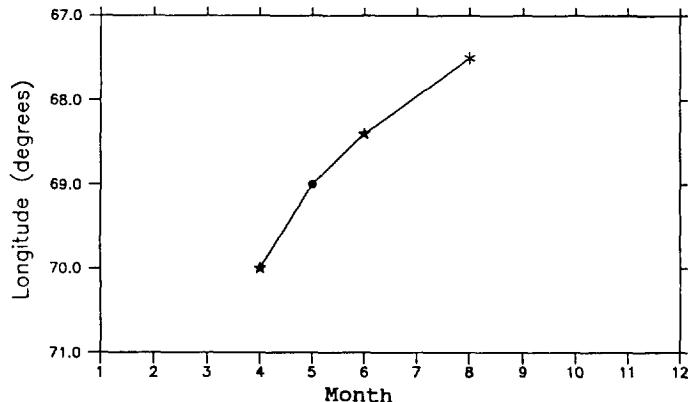


Fig. 5. Monthly mean position of the eastern boundary of the surface (2 m) low-salinity plume defined by the 32.0 contour. They are estimated from different hydrographic data sets taken during 26–29 April 1988, 6–12 June 1989 (★), 11–21 May 1976 (●, LIMEBURNER and BEARDSLEY, 1982) and 28 July–5 August 1987 (*, GARRISON and BROWN, 1989).

found that the predicted frontal position for tidal plus wind mixing shifted to $\log_{10}(h/(D + 0.59)) = 1.65$, and the predicted well-mixed area is slightly expanded from the case of tidal mixing alone (see the 1.65 dash-dot line in Fig. 9). On the other hand, while the oscillatory advection and monthly and fortnightly variation in the strength of tidal mixing cause a 10-km deviation from the $\log_{10}h/D = 1.9$ contour, the M_2 – S_2 spring–neap cycle is sufficiently large that the transition zone between the well-mixed and stratified regions will shift to $\log_{10}h/D = 2.1$ during the spring tide when the tidal dissipation is larger.

The June 1989 CTD survey showed a continuous contour of $\Delta\sigma_t = 0.5$ (the boundary between the well-mixed and stratified regions defined by GARRETT *et al.*, 1978) from Nantucket Shoals to the western flank of Georges Bank between the 80- and 100-m isobaths, which was in good agreement with the mean position of the SST front obtained from daily satellite SST images for the period 5 May to 12 June 1989. This front position was identified at two SCOPEX small-scale biological survey stations on both sides of the northern GSC made on 26 May and 3 June 1989 (Fig. 9). Large temperature and σ_t gradients were observed west of the GSC on 26 May 1989 at CTD Stas 85 and 86 (see LIMEBURNER and BEARDSLEY, 1989). CTD Sta. 85 showed a relatively well-mixed water column with a vertical difference of 0.4°C in temperature and 0.1 in σ_t in the upper 40 m, while CTD Sta. 86 showed a strong thermocline from 10 to 25 m near the surface. Using continuous SST data obtained between these two stations, we found that a thermal front with a sharp jump in temperature of 1.4°C over a distance of 0.4 km was located at 69°21.30'W and 41°23.75'N, over Nantucket Shoals where the water depth is about 61 m. A similar thermal front with a temperature jump of 2.0°C over 1.0 km at the surface was also observed east of the GSC (at 68°35.15'W and 41°23.85'N) on 3 June 1989 near CTD Sta. 136, where the water depth is about 80 m. A similar front structure was observed in April 1988, implying that a relatively permanent vertical mixed region existed in the GSC near the 80-m isobath and the vertical mixed zones over Nantucket Shoals and Georges Bank were connected. This observed result is significantly different from those predicted by GARRETT *et al.* (1978) for the case of tidal mixing alone and LODER and GREENBERG (1986) for the case of tidal plus wind mixing but coincides well with the continuous contour

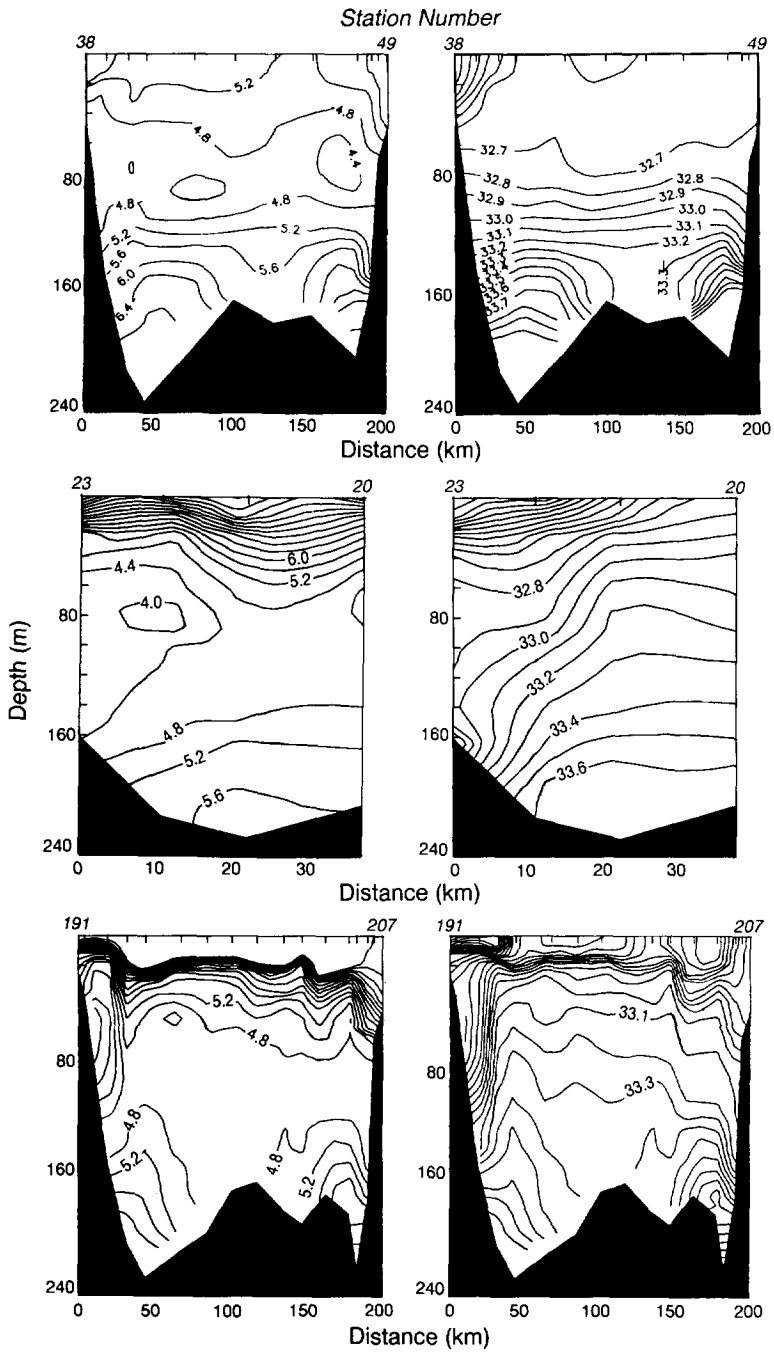


Fig. 6. Distributions of temperature (left) and salinity (right) on transect E of the 26–29 April 1988 survey (top), transect F of the 18–21 May 1989 survey (middle), and transect E of 6–12 June 1989 survey (bottom). These are the northernmost transect shown in Fig. 1.

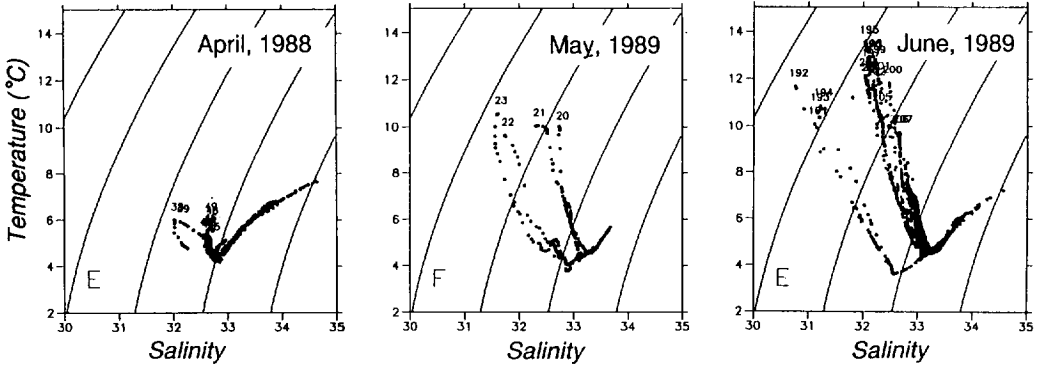


Fig. 7. *T-S* diagrams for transect E of the 26–29 April 1988 survey (left), transect F of the 18–21 May 1989 survey (middle), and transect E of the 6–12 June 1989 survey (right). The numbers in the diagrams are the CTD station identifiers, and the curves are the contours of constant σ_t from 23.0 (left) to 27.0 (right).

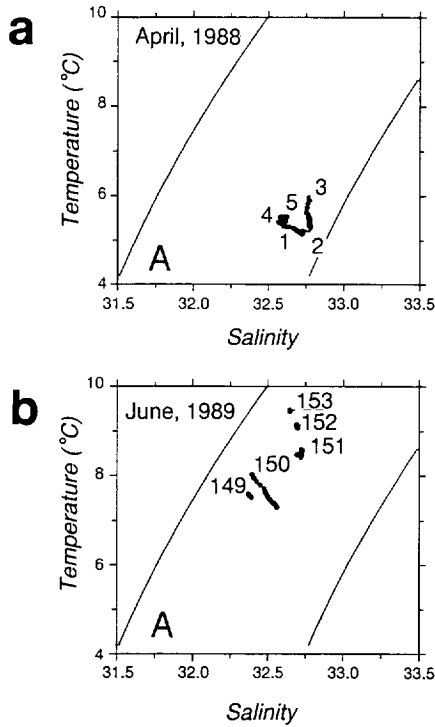


Fig. 8. *T-S* diagrams for transect A of the (a) 26–29 April 1988 and (b) 6–12 June 1989 surveys. The numbers in the diagrams are the CTD station identifiers, and the curves are the contours of constant σ_t from 23.0 (left) to 27.0 (right).

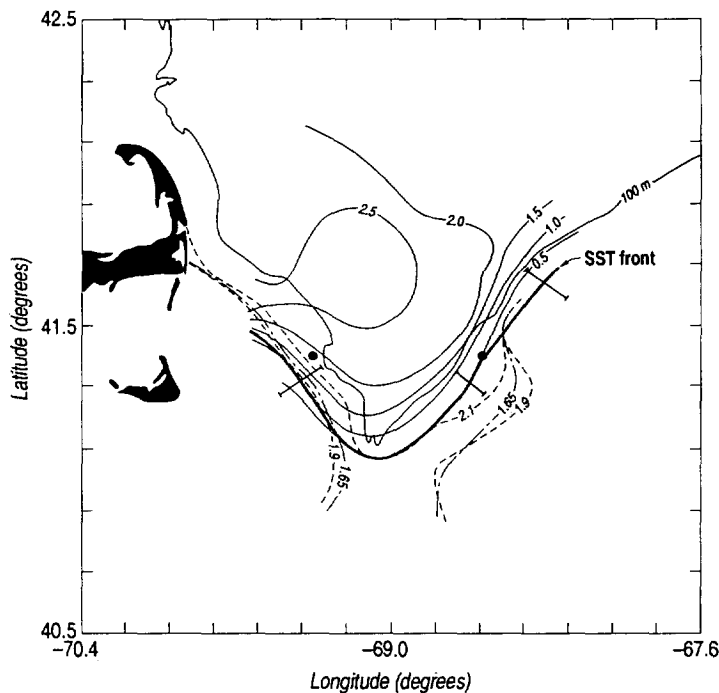


Fig. 9. Comparison between observed and model tidal fronts over Nantucket Shoal, GSC, and Georges Bank. Solid lines are $\Delta\sigma_t$ contours in the upper 40 m calculated from the June 1989 CTD survey data. Heavy solid line is the mean position of the thermal front averaged over satellite SST images from 5 May to 12 June 1989. The deviation of the SST thermal front to the mean position is indicated by the cross front solid line. ● is the frontal position found during the small-scale biological surveys made on 26 May and 3 June 1989. Dashed and dash-dot lines are the contours of $\log_{10}h/D = 1.9$ and 2.1 for the case of tidal mixing alone, and $\log_{10}(h/(D + 0.59)) = 1.65$ for the case of tidal plus wind mixing, respectively. They are digitized from Figs 3 and 5 in LODER and GREENBERG's (1986) paper.

of $\log_{10}h/D = 2.1$ across the center of the GSC estimated by LODER and GREENBERG (1986) for the spring tide (see Fig. 9).^{*} This suggests that tidal mixing in the GSC during late spring is more intense than that predicted by the barotropic M_2 tidal dissipation and wind mixing alone.

5. MECHANISMS FOR THE OFFSHORE SPREADING OF THE LOW-SALINITY PLUME

In general, a buoyancy-driven density current due to river discharge flows like a boundary jet along the coast in the direction of the propagation of coastal trapped waves.

^{*}Note: Since spatial averaging was done to smooth the water depth in GREENBERG's (1979) tidal model, the positions of the 50-m and 100-m isobaths LODER and GREENBERG (1986) refer to are a little different from the real water depth, especially on the northern flank of Georges Bank where large gradients of bottom topography exist. For this reason, we plot these three contours referred to the real 100-m isobath rather than a direct copy of their absolute positions from Figs 3 and 5 in LODER and GREENBERG's (1986) paper.

In the absence of sloping bottom topography over the shelf, the low-salinity plume may turn seaward from the coast due to a downstream accumulation of water caused by convergent currents along the coast (STERN *et al.*, 1982; CHAO, 1988). Whenever a sloping bottom is present to impose geostrophic contours parallel to the coast, the plume tends to be trapped near the coast so that there is no offshore spreading of the plume in the downstream direction (CHAO, 1988). An increasing river discharge in late spring should only intensify the strength of a buoyancy-driven coastal current, and not lead directly to offshore spreading. Thus, the observed spreading of the low-salinity plume east of Cape Cod in June 1989 implies some additional dynamical mechanism.

Two additional mechanisms involve local forcing by the along-coast component of wind stress and deeper circulation. Since the time scale on which the low-salinity plume evolved was much greater than inertial, and water within the plume was relatively well mixed and almost uniform along the coast, the contribution of gradients of sea-surface elevation and baroclinic pressure to the offshore transport can be neglected. Without considering non-linear advection, the along-coast vertically averaged momentum equation in the plume can be thus simplified to

$$\rho f h \bar{u} = \tau_s + \tau_d \quad (1)$$

where \bar{u} is the vertically averaged offshore velocity in the plume, h the mean plume thickness, f the Coriolis parameter, ρ the water density, τ_s the along-coast surface wind stress, and τ_d the along-coast interfacial stress at the base of the plume *forced by the deeper circulation*. This balance suggests that the offshore transport of the low-salinity plume is determined by the stresses due to both surface wind and the deeper current.

The time scale of the response of the GOM to both local and remote wind forcing is about 2–20 days (BROWN and IRISH, 1992). East of Cape Cod, a northward upwelling-favorable wind should tend to cause a seaward spreading of the low-salinity plume by the offshore Ekman transport and cause strong vertical stratification offshore through the resulting upwelling flow. In contrast, a southward downwelling-favorable wind should tend to confine the low-salinity plume to the coast by onshore Ekman transport and cause the plume to deepen in a narrow band near the coast by the downwelling flow. A strong downwelling-favorable wind higher than 15.0 m s^{-1} was observed 14–20 days before the April 1988 survey, while a moderate and steady upwelling-favorable wind with an average speed of about 5 m s^{-1} occurred frequently for about one month before the June 1989 survey (Fig. 10). The structure of the low-salinity plume observed in late spring of 1988 and 1989 is qualitatively consistent with this idea of local wind forcing.

The offshore Ekman transport in late spring 1989 can be estimated using the wind data from the NOAA meteorological buoy at 69.6°W and 40.5°N . Based on the neutral, steady-state drag coefficient developed by LARGE and POND (1981), the resulting wind stress for a 5 m s^{-1} wind speed measured at the height of 13.8 m is $\tau_s = 0.34 \text{ dyn cm}^{-2}$, and hence the resulting eastward offshore Ekman transport is $U_E = \tau_s / \rho f = 0.35 \text{ m}^2 \text{ s}^{-1}$ at 41°N . Taking the 32.0 contour as the boundary of the low-salinity plume, the total volume of the plume increased from about $0.7 \times 10^{10} \text{ m}^3$ in 18–21 May 1989 to about $1.2 \times 10^{11} \text{ m}^3$ in 6–12 June 1989, finally covering an area of 0.8° latitude \times 0.8° longitude with a thickness of 20 m . If the plume was pushed offshore only by Ekman transport with a value of $0.35 \text{ m}^2 \text{ s}^{-1}$ from late May to mid-June 1989, it would take about 44 days to form the plume observed in the mid-June survey. Since this is longer than the 20-day time interval between the May and June 1989 surveys, we conclude that the offshore spreading of the low-salinity surface

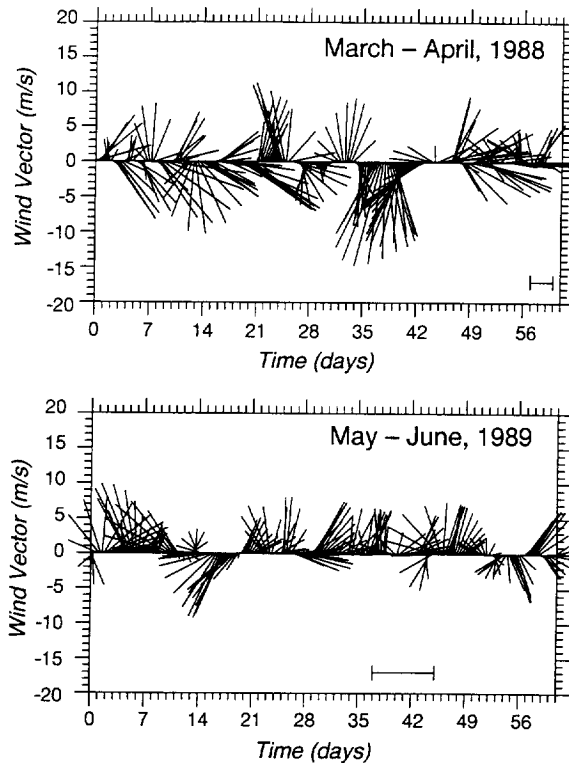


Fig. 10. The 40-h, low-passed wind vectors measured at 13.8 m on the Nantucket Light NOAA environmental buoy No. 44008 (69.5°W, 40.5°N) from 1 March to 30 April 1988 (upper) and 1 May to 30 June 1989 (lower). The vectors are subsampled with a 6-h interval, and horizontal brackets indicate the period of the main regional CTD/ADCP surveys.

plume in late spring 1989 may also be driven by the internal stress due to the deeper circulation.

There is a relatively permanent cyclonic subtidal circulation in the intermediate-depth water below 50 m in the northern GSC, flowing southeastward along the 100-m isobath east of Cape Cod and then turning northeast to join the anti-cyclonic gyre around Georges Bank (CHEN *et al.*, 1995: Fig. 11). This flow in the upper MIW may produce a stress on the low-salinity surface plume so as to help drive the upper-layer water offshore from Cape Cod. The directly measured vertically averaged eastward residual velocity in the western side of the GSC was about 3–4 cm s^{-1} in the low-salinity surface plume in June 1989, which caused an eastward transport of 0.6–0.8 $\text{m}^2 \text{s}^{-1}$ for the plume with a thickness of 20 m, about twice the mean wind-driven Ekman transport computed from May–June 1989. This suggests that the eastward transport due to the interfacial stress was about 0.25–0.45 $\text{m}^2 \text{s}^{-1}$, which is roughly comparable to the wind-driven Ekman transport. Therefore, we suggest that the offshore spreading of the low-salinity surface plume in the northern GSC during spring was driven by both the interfacial stress due to the deeper cyclonic circulation and local upwelling-favorable wind stress.

6. CONCLUSIONS

The surface salinity patterns observed in late April 1988 and May–June 1989 differ significantly in the extent of the freshwater plume which occurs east of Cape Cod in spring. In April 1988, the surface plume was just beginning to form along the outer coast of Cape Cod while 3–6 weeks later in 1989, the minimum salinity was about 1.5 less and a large pool of water fresher than 31.6 had pushed eastward over much of the northern GSC region. The difference in the amount of freshening observed between the two surveys is due primarily to the 6-week difference in the seasonal cycle and increased river discharge in 1989. The seasonal offshore extension of the low-salinity plume may be driven by both an upwelling-favorable wind stress and a deeper cyclonic circulation.

A notable difference was found at mid-depth in the structure of MIW in late spring during the 1988 and 1989 surveys. In April 1988, the seasonal pycnocline was just beginning to form, and the spatial structure of MIW was relatively uniform. In May–June 1989, a narrow tongue of temperature minimum water was found along the western flank of the northern GSC between 40 and 120 m. The minimum temperature and salinity in the core of this tongue became colder and fresher with time between May and June. Mixing between this colder and fresher water and the interior MIW occurred as the tongue flowed southward into the northern GSC. Unfortunately, the SCOPEX measurements do not clearly identify the upstream source region of this temperature-minimum water, nor why this narrow tongue of temperature minimum water was observed in late spring 1989 but not earlier in the seasonal cycle in 1988. What is the source of this temperature-minimum water, the unusually large river discharge in the western GOM which occurred in 1989, or a difference in the surface heat flux during the preceding winter, or the seasonal evolution of a coastal-trapped buoyancy-driven current that forms upstream each year and flows into the northern GSC in May? Future field work including hydrographic measurements with high spatial resolution in the cross-isobath direction will be needed to better define this new and unexpected water mass feature.

A relatively permanent continuous thermal front separating well mixed water from stratified water was found along the eastern flank of Nantucket Shoals, across the northern shallow region of the GSC and along the northwestern flank of Georges Bank. The location of this front moved roughly 10 km in the north–south direction, most likely due to oscillatory tidal advection and monthly/fortnightly variation in the strength of tidal mixing in the GSC. Comparison of the location of this front with theoretical predictions by LODER and GREENBERG (1986) suggests that enhanced tidal mixing due to the spring-neap cycle is important in determining the relative balance between buoyancy input and tidal mixing in the GSC region.

Acknowledgements—This research was supported by the National Science Foundation under grants OCE 87-13988 and OCE 91-01034. We would like to thank Ken Brink, Dave Chapman, Glenn Flierl, Glen Gawarkiewicz, Steve Lentz and Paola Rizzoli for their interest in this work and valuable comments and suggestions. We also want to thank Jack Jossi, who permitted us to use the surface salinity data obtained on the NOAA Ship Of Opportunity Program (SOOP), Howard Winn and Karen Wishner who asked us to participate in SCOPEX, and Peter Cornillon, who provided the satellite sea-surface temperature data.

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