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The annual variation of water mass structure in the Gulf of Maine: 1986–1987

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ABSTRACT

The annual variation in the structure and disposition of the principal water masses in the Gulf of Maine has been investigated with a set of water property observations using five shipboard surveys and four moored arrays with data telemetry. The time series observations document the cooling-induced destratification of the upper water column during autumn and the subsequent mixed-layer deepening in the western Gulf—primarily in Wilkinson Basin—during winter 1987. Unusually large amounts of fresher Scotian Shelf Water inflow inhibited winter 1987 vertical mixing in the eastern Gulf relative to the deep mixing in and around Wilkinson Basin in the western Gulf. The net result of these processes was a 1987 Gulf that was colder and fresher than the 1986 Gulf. Detailed histories of the thicknesses of the principal water masses in the Gulf—namely, Maine Surface Water, Maine Intermediate Water, and Maine Bottom Water—at the mooring sites reveal the early summer progression of Slope and Bottom Water from the Northeast Channel to Georges Basin and on to Jordan Basin. Maine Intermediate Water made up nearly 50% of the entire volume of the Gulf in early spring 1987. During the summer, Maine Intermediate Water in the eastern Gulf was replaced by a warmer water mass we call Summer Intermediate Water. A spring-summer 1987 sequence of CTD-derived water mass distribution maps documents (1) the retreat of Maine Intermediate Water into Wilkinson Basin, (2) the westward spread of Summer Intermediate Water, and (3) the inflow of Slope Water. A simple water mass conservation model indicates that 72% of the Maine Intermediate Water loss flows out of the observation domain at a rate of 0.21×10^6 m^3/s, while the other 28% contributes, through mixing with Surface Water and Bottom Water in the eastern Gulf, to the production of Summer Intermediate Water. The combined inflow of Slope Water (0.11×10^6 m^3/s), Bottom Water (0.03×10^6 m^3/s), and Summer Intermediate Water (0.07×10^6 m^3/s) appears to have balanced the April-July outflow of Maine Intermediate Water.

1. Introduction

A significant part of the non-tidal circulation in the Gulf of Maine is related to its evolving density structure (Bigelow, 1927; Brooks, 1985; Brown and Irish, 1992). The density and water property structure of the Gulf is primarily controlled by (a) the

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inflow of fresher, near-surface water from the Scotian shelf, and saltier, deeper water from the upper continental slope, (b) the modifications of water properties by mixing, winter cooling, and local fresh water runoff, and (c) the outflow of surface and intermediate water to Georges Bank and the adjacent shelf. Hopkins and Garfield (1979) have defined the water mass structure in the Gulf in terms of Maine Surface Water, Maine Intermediate Water and Maine Bottom Water and generally described their annual evolution.

According to Hopkins and Garfield (1979), Maine Surface Water characteristics vary widely and are controlled by seasonal surface cooling and warming within the Gulf, as well as inflow of Scotian Shelf Water and local fresh water runoff. Scotian Shelf Water provides most of the fresh water to the Gulf. The freshness of Scotian Shelf Water is due to contributions from the St. Lawrence River (Sutcliffe *et al.*, 1976) and remote Arctic regions (Fairbanks, 1982; Chapman and Beardsley, 1989). Once in the Gulf, Scotian Shelf Water mixes to depths of at least 150 m in some parts of the Gulf (Chapman *et al.*, 1986) during the winter.

Maine Intermediate Water is a remnant of that water mass produced during the previous winter. Thus the properties of Maine Intermediate Water, most notably its very cold temperatures and freshness, are redefined each year by the relative amounts of residual Maine Intermediate Water from the previous year, the fall/winter Scotian Shelf Water inflow, as well as the intensity of wintertime overturning and thus entrainment of Bottom Water. Each spring, vernal warming isolates that year's Maine Intermediate Water from the near-surface Maine Surface Water (Hopkins and Garfield, 1979). During the summer, the layer of cold Maine Intermediate Water thins and fragments, particularly in the eastern Gulf, due to export from the Gulf through the Northeast Channel and mixing with the other Gulf water masses. Some of the export from southern Wilkinson Basin is associated with an intense northeastward flowing current located along the north flank of Georges Bank—the north flank jet (Magnell *et al.*, 1980). According to Hopkins and Garfield (1981), some of the Wilkinson Basin Maine Intermediate Water in the north flank jet leaks onto Georges Bank. The work of Flagg (1987) suggests that the north flank jet exits the Gulf on the south side of the Northeast Channel and is connected to a cold band of water which flows southwestward along the 80 m isobath on the south flank of Georges Bank. He presents evidence that this cold band of water on southern Georges Bank flows onto the New England shelf and into the Mid-Atlantic Bight region.

Maine Bottom Water is derived from Slope Water which flows at depth into the Gulf through the Northeast Channel (sill depth is 232 m). According to Gatien (1975), the Slope Water inflow alternates between water with Labrador Current origins and a slightly warmer, saltier counterpart with Gulf Stream origins. Slope Water mixes with Maine Intermediate (and perhaps some cold Maine Surface Water) to produce Maine Bottom Water. The movement of Maine Bottom Water is controlled by the rugged bathymetry associated with Wilkinson, Georges and Jordan Basins. That bathymetry is characterized by depths greater than 270 m and interbasin sill depths in the 180–190 m range. As Maine Bottom Water flows through the Gulf, it is modified further through mixing primarily with the colder, fresher Maine Intermediate Water. Mountain and Jessen (1986) present hydrographic evidence that suggests that Maine Bottom Water can become fresher and colder as it flows from Georges Basin preferentially through Jordan Basin to Wilkinson Basin.

The purpose of this paper is to use a comprehensive set of moored and ship survey temperature/conductivity observations from the period August 1986 through September 1987 to document the details of the annual variation in the structure and volume distributions of the Gulf of Maine water masses. In Section 2 the field observation program is described. In Section 3 the seasonal variability of water properties and their importance to water mass distribution are described. In Section 4 the changes in the structure and volume distributions of the spring/summer 1987 Gulf water masses are presented. The implications of water mass volume change for mixing and transport in the Gulf are discussed in terms of a simple water mass conservation model. In Section 5, the spring and summer water mass histories for the 1987 Gulf are presented. A summary of the principal results appears in Section 6.

2. 1986–1987 field program description

During 1986 and 1987, the University of New Hampshire obtained moored and shipboard measurements to describe the yearly cycle of water mass formation and erosion in the Gulf of Maine (Fig. 1). Our goal was to measure the variability in the temperature, salinity, density and pressure fields and to determine their effects on circulation in the Gulf. Spatial distributions of hydrographic information were obtained during Gulf-wide surveys on the RV *Gyre* during 3–16 August 1986 (Garrison and Brown, 1987), the RV *Oceanus* during 5–15 February 1987 (Moody *et al.*, 1990), RV *Endeavor* during 4–18 April 1987 (Garrison *et al.*, 1989), 28 July–8 August 1987 (Garrison and Brown, 1989a), and 3–16 September 1987 (Garrison and Brown, 1989b). The typical survey pattern (Fig. 2) was designed to map the hydrography of the water of the Gulf with an emphasis on the basins and water deeper than 100 m. Temporal variability of the Gulf hydrography was measured through deployments of moored temperature, conductivity, and bottom pressure instrumentation in each of the three major basins of the Gulf—namely, Wilkinson, Jordan and Georges—as well as the Northeast Channel (Fig. 1). Moored arrays, like that depicted in Figure 3, monitored the time variability of temperature and conductivity (and derived salinity and density) at 50 m depth intervals at the four sites for the time periods indicated in Figure 4 (Table 1).

a. Moored array instrumentation. Temperature/conductivity observations were made using surface moored arrays, which provided near-realtime data to our laboratory.

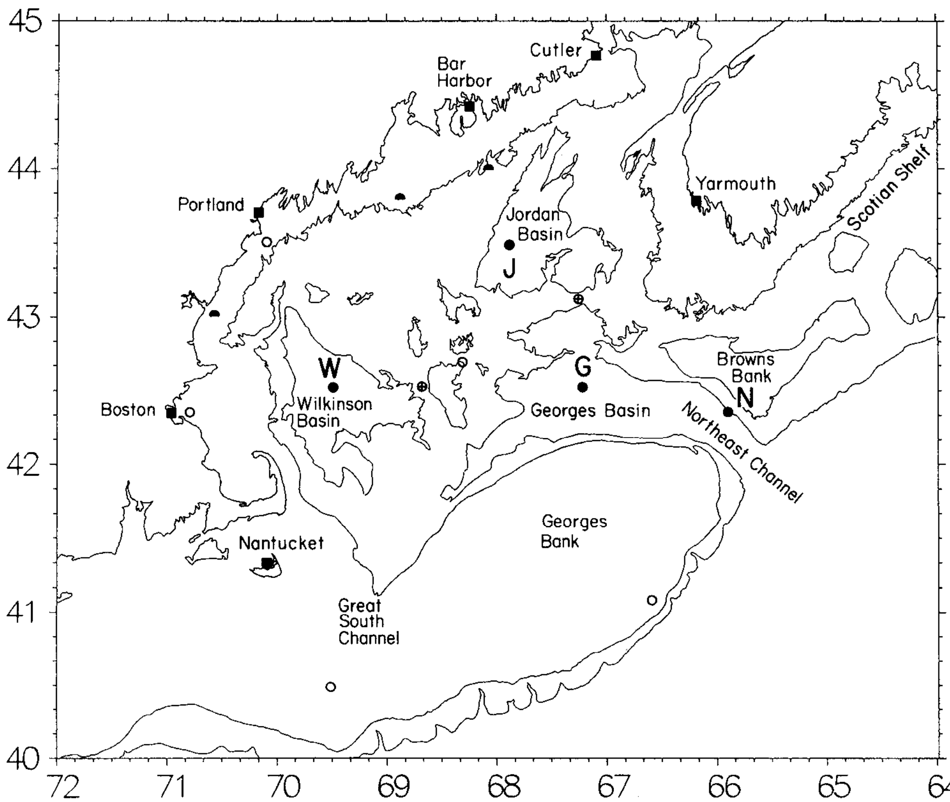


Figure 1. Location map of the Gulf of Maine. The 100 m and 200 m isobaths define the relatively shallow Georges and Browns Banks as well as the three main basins. Typical sill depths between basins are 180 m. The temperature/conductivity/bottom pressure instrumentation (solid circles); federal U.S. and Canadian sea level stations (solid squares), National Data Buoy Center (NDBC) meterological buoys (open circles), NDBC island C-MAN meteorological stations (solid half circles), and two USGS long-term near bottom current meter sites (crossed circle) are also located.

The surface buoy (Fig. 3) consisted of a flotation sphere with damping plate below and a tower above. The tower held a satellite antenna, guard lamp, radar reflector, air sensors, and the electronics package. Solar panels supplied power for the sensors, data system and satellite transmitter. The panels were mounted low on the buoy so that they were continually cleaned by the waves. The surface buoy was moored to the bottom with a composite electro-mechanical cable. To increase mooring life, in-line compliant members were used to reduce wear of the mooring components due to surface wave action. The overall buoy system performed well, with only one instance of suspected elastic tether failure. The mooring hardware showed little if any mechanical wear after 13 months (see Wood and Irish, 1987, for further detail).

The heart of the buoy data acquisition and telemetry system was a Synergetics data collecting platform, a microprocessor-controlled system to which a Sea Data cassette

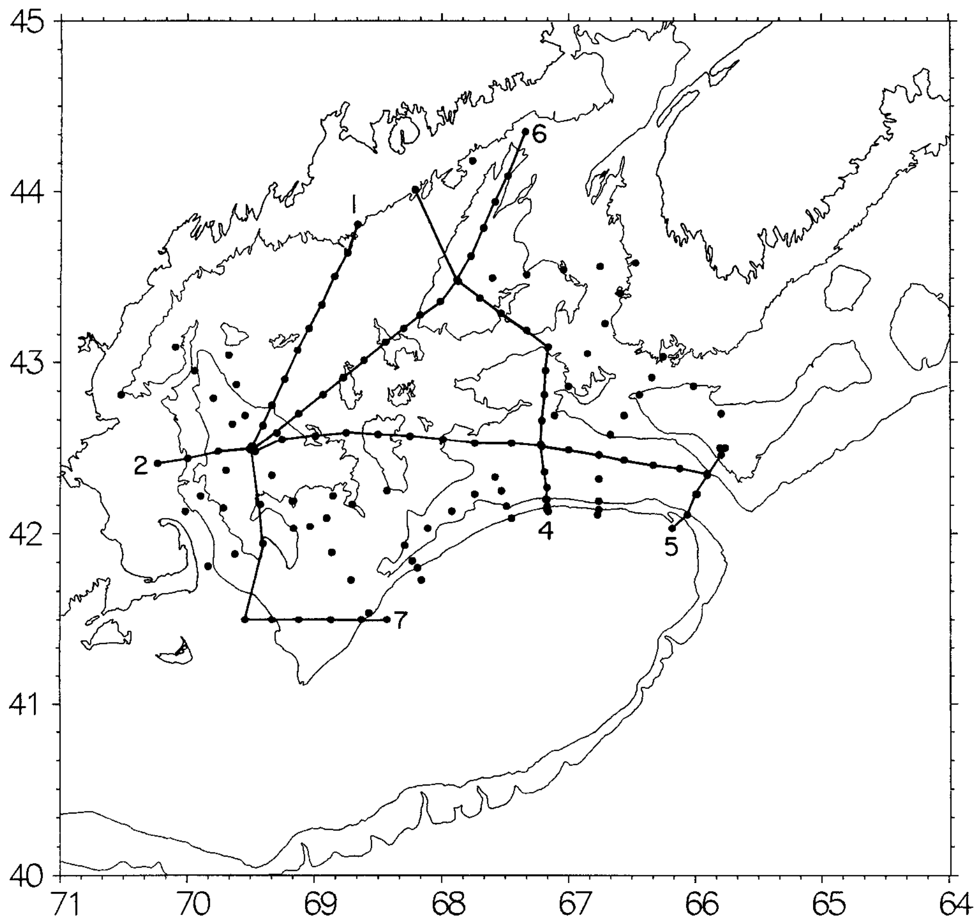
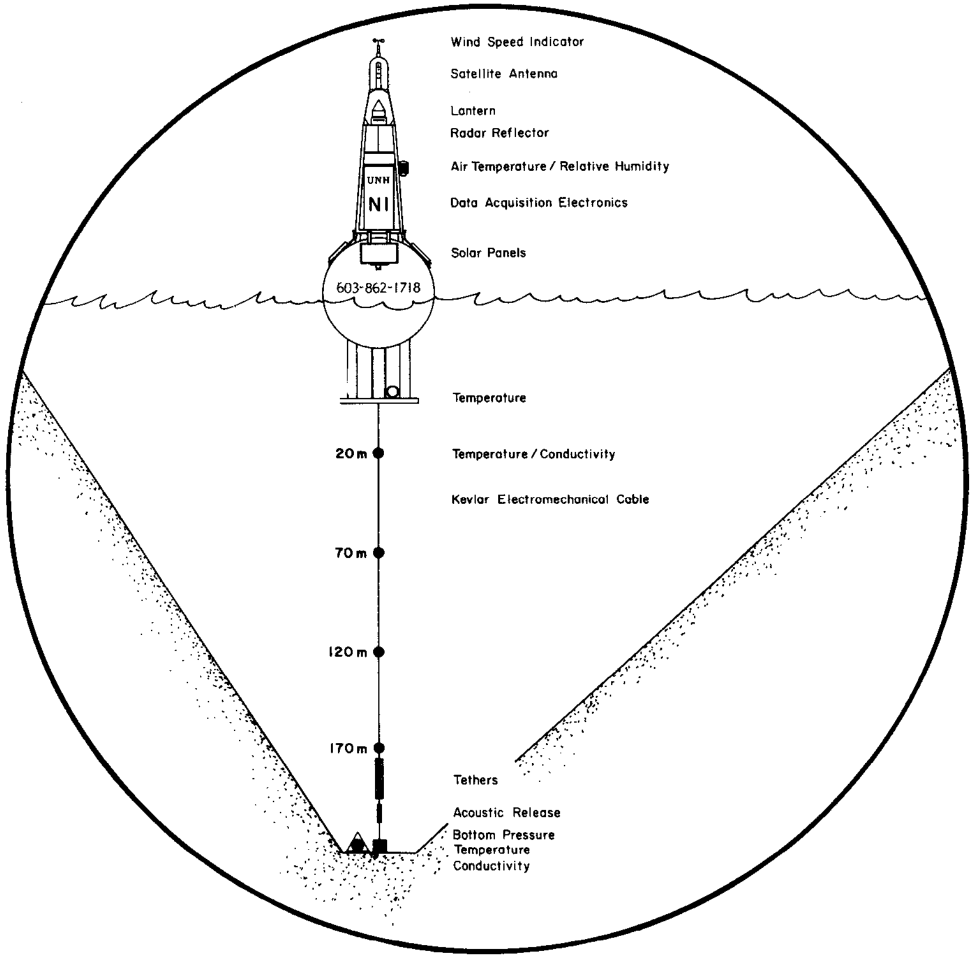


Figure 2. Typical basin-centered CTD survey coverage during the Gulf of Maine 1986–87 study. A selected set of vertical sections are also identified.

recorder was added for backup (see Irish *et al.,* 1987, for details). The system was programmed to control the power, sampling and telemetry process. A precision quartz crystal clock controlled the timing of the sampling and data telemetry to the GOES satellite. An ARGOS transmitter was added to the Northeast Channel buoy primarily to track the buoy if it were to break loose. The ARGOS system also returned some data and diagnostic information on system status. The system controlled power to all sensors and digitized the voltages and frequencies approximately every minute.

Every hour the averaged samples and present system status were written to the cassette tape recorder and the ARGOS transmitter. The hourly data for each sensor were also stored in a six-sample buffer, along with information on the system status, the battery voltages and the system temperature. Every three hours, the last six hours of data were transmitted ashore by GOES to the National Environmental Satellite



Northeast Channel T/C Array

Figure 3. Typical temperature/conductivity/bottom pressure instrumentation array for the Gulf of Maine 1986–87 observations. A surface data acquisition and telemetry buoy is hard-wired to an array of temperature/conductivity (T/C) sensor pairs. The T/C array is connected to its anchor via a set of elastic tethers to absorb surface water action. A separate bottom pressure/conductivity/temperature instrument is deployed aside the T/C array.

and Data Information Service (NESDIS) of NOAA. This was a 100% redundant transmission scheme which permitted recovery of data lost from a single missed transmission.

This quasi-realtime data were retrieved daily at UNH. An automated program accessed the data at NESDIS through a dialup telephone link and controlled its transmission to UNH. A user-initiated suite of programs at UNH was used to check

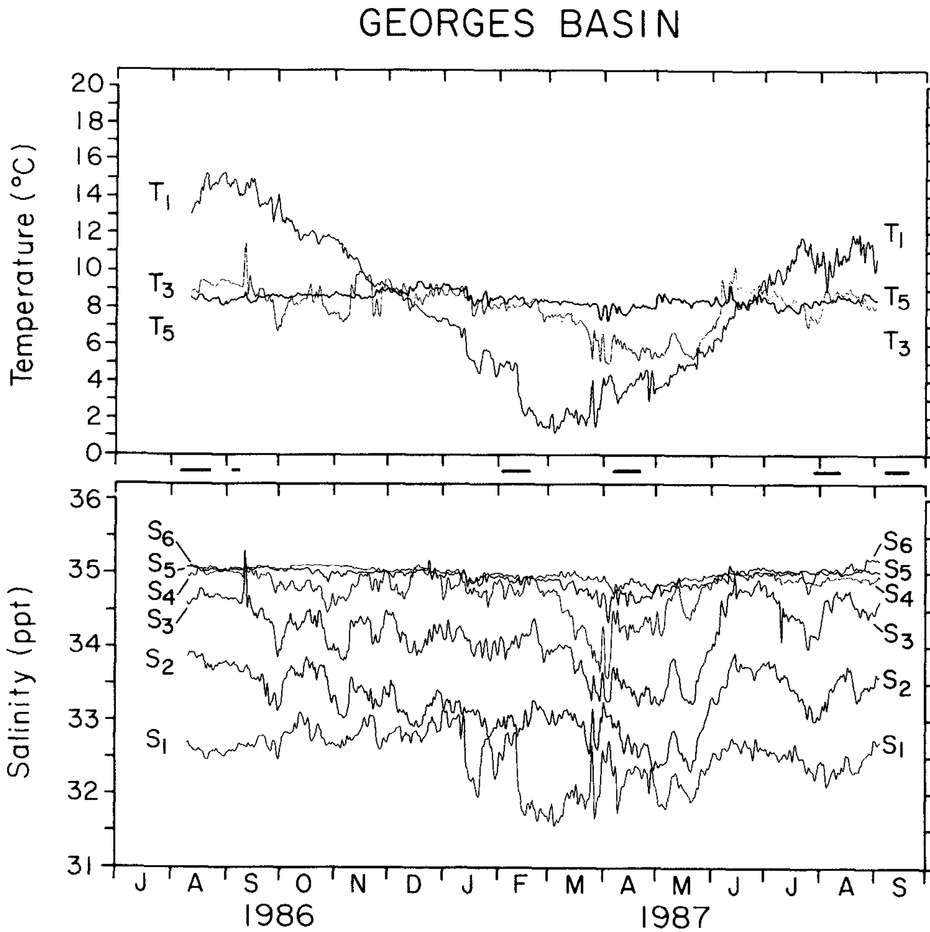


Figure 4. Subtidal moored temperature and salinity time series in Georges Basin (see Table 1). The annual cooling/warming cycle is most obvious at shallower depths represented by T_1 (21 m). Some seasonality is seen at mid-depth temperature represented by T_3 (120 m). The temperatures of the deeper waters represented by T_5 (220 m) are relatively steadier. The time periods of the hydrographic cruises are indicated.

redundant data, check time continuity, normalize data and create the series files for further processing.

Hourly averages of water column temperature and conductivity were measured at the depths listed in Table 1 with Sea Bird sensors (Pederson, 1969; Pederson and Gregg, 1979), which were hard-wired to the data acquisition system on the surface buoy. Tributyltin cells were attached to the conductivity sensors to minimize biofouling.

Fifteen-minute averages of bottom pressure/temperature/conductivity were mea-

Table 1. Gulf of Maine moored temperature (T), conductivity (C), and bottom pressure (P) measurement summary, August 1986 through September 1987. The standard deviations for the hourly and subtidal temperatures, salinities (instead of conductivity), and pressures are given in °C, psu, and decibars, respectively.

Mooring location	Sensor ID	Depth (m)	Start time End time	Standard deviations Total	Subtidal
Georges Basin	T0	1	8 Aug 86 1800	4.83	4.79
42° 31.4′ N	T1	21	5 Sep 87 1400	3.99	3.93
67° 13.0′ W	C1			0.37	0.35
	T2	71		1.89	1.83
	C2			0.44	0.39
	T3	120		1.27	1.21
	C3			0.48	0.43
	T4	170		0.73	0.69
	C4			0.30	0.28
	T5	220		0.36	0.33
	C5			0.13	0.12
	T6	269		0.45	0.39
	C6			0.09	0.08
	T7	328	10 Aug 86 0700	0.43	0.43
	C7			0.38	0.43
	P1		5 Sep 87 0700	0.60	0.06
	P2			0.60	0.06
Jordan Basin	T0	1	8 Aug 86 1600	3.74	3.69
first	T1	19	13 Feb 87 0300	2.69	2.64
deployment	C1			0.31	0.31
43° 29.6′ N	T3	118		0.65	0.60
67° 52.6′ W	C3			0.19	0.17
	T4	169		0.33	0.27
	C4		(failed 6 Nov 86)	0.13	0.10
	T5	218	(failed 14 Sep 86)	0.20	0.15
	C5		(failed 31 Aug 86)	0.01	0.01
	T6	285	7 Aug 86 0200	0.47	0.46
	C6			0.26	0.26
	P1		10 Sep 87 1400	0.96	0.08
	P2			0.96	0.07
second	T0	1	19 Apr 87 1800	4.07	3.99
deployment	T1	16	6 Sep 87 1500	2.75	2.61
43° 29.3′ N	C1			0.27	0.26
67° 52.5′ W	T2	66		1.59	1.51
	C2			0.35	0.34
	T3	115		1.14	1.09
	C3			0.41	0.40
	T4	165		0.99	0.96
	C4			0.46	0.45
	T5	217		0.55	0.55
	C5			0.31	0.30

Table 1. (Continued)

Mooring location	Sensor ID	Depth (m)	Start time End time	Standard deviations Total	Subtidal
Northeast	T0	1	3 Sep 86 0000	2.80	2.71
Channel	T1	21	4 Feb 87 0600	3.91	3.83
first	C1			0.87	0.78
deployment	T2	72		2.00	1.87
42° 21.0′ N	C2			0.85	0.71
65° 54.2′ W	T3	116	(failed 29 Nov 86)	1.52	1.38
	C3		(failed 29 Nov 86)	2.43	0.37
	T4	161		1.15	1.01
	C4			0.67	0.49
second	T0	1	5 Apr 87 2000	4.76	4.71
deployment	T1	21	4 Sep 87 1800	3.27	2.97
42° 21.0′ N	C1			0.43	0.40
65° 54.2′ W	T2	72		1.79	1.65
	C2			0.53	0.46
	T3	116		1.55	1.30
	C3			0.50	0.40
	T4	161		0.94	0.78
	C4			0.27	0.21
	P, T, C			(Instrument not covered)	
Wilkinson	T0	1	14 Jan 87 1400	1.10	1.09
Basin	T1	16	7 Apr 87 1200	1.11	1.10
42° 30.8′ N	C1			0.27	0.27
69° 29.0′ W	T2	66		0.93	0.89
	C2			0.17	0.17
	T3	115		0.69	0.65
	C3			0.11	0.11
	T4	165		0.42	0.41
	C4			0.15	0.15
	T6	273		0.04	0.04
	T7			0.13	0.13
	P1			0.82	0.10
	P2			0.82	0.10

sured with bottom instrumentation deployed next to each moored temperature/conductivity mooring. The bottom instruments, similar to those discussed by Brown *et al.* (1985), had dual Paroscientific pressure sensors (Wearn and Larsen, 1982), a Sea Data end cap thermistor, and a Sea Bird conductivity sensor.

The temperature sensors were calibrated before deployment, and the stability was checked through comparison with CTD profiles. No measurable temperature drifts were observed. Temperature accuracies were within ±0.01°C with resolution better than 1 millidegree Centigrade. While the conductivity sensors were calibrated both before and after deployment, comparisons with CTD measurements showed that these calibrations could not be used to document drift. In the case of the conductivity measurement away from the bottom, biofouling would lead to a gradual drifts as large as 0.02 S/m/yr (0.2 psu/yr). These were corrected through comparisons with CTD measurements. In the case of bottom conductivity measurements, suspended sediment in the water column settling in the sensor cell resulted in both intermittent spikes or drift toward lower conductivity (salinity). These spurious data were identified and corrected by comparing computed salinities and densities with those on the nearby moorings. Corrections for the secular drift was more difficult to determine because high currents would occasionally sweep the sediment out of the sensor cell. Thus, we could not correct many of the bottom conductivities well enough to use them in this study.

b. Water property measurements. The temperature/conductivity mooring in Georges Basin had a near-surface water temperature sensor and six temperature/conductivity sensor pairs located at 50 m intervals between 20 m and 270 m (see Table 1). The moored instruments returned nearly thirteen months of error-free records from all sensors except for the bottom conductivity record which was terminated in May due to sediment contamination. The hourly observations were filtered using a lowpass filter, with a $[36 \text{ hour}]^{-1}$ cutoff frequency, to remove tidal and high frequency fluctuations. The resulting "subtidal" time series of temperature and salinity for the Georges Basin mooring are presented in Figure 4. The annual warming/cooling cycle is most evident in the measurements in the upper 120 m. Note that the water at 20 m is both the warmest in the summer and coldest in the winter. However, in Georges Basin, the lowered winter temperatures are compensated for by reduced salinities. In fact, there is evidence for a significant annual cycle in salinity all the way down to depths of 220 m.

The temperature/conductivity mooring in Jordan Basin had a near-surface water temperature sensor and five temperature/conductivity sensor pairs located at 50 m intervals between 20 m and 220 m (Table 1). Several sensor failures from undetermined causes marred the first half data return. To add insult to injury, the Jordan Basin array broke loose on 11 February 1987 during an intense storm, probably due to elastic tether failure. The buoy, with cable and sensors, drifted out of the Gulf into the Gulf Stream, and was carried across the Atlantic and subsequently beached near Brest, France one year to the day later! The telemetry scheme had indeed demonstrated its value. Despite the mooring loss, we acquired seven months of useful data from the mooring before it broke loose—data which could otherwise have been permanently lost if the data acquisition canister had not been recovered.

The original Jordan Basin array was replaced on 19 April 1987 with a similar array which had been deployed previously in Wilkinson Basin between January and April 1987. The Wilkinson Basin array had a near-surface water temperature sensor and five temperature/conductivity sensor pairs located at 50 m intervals between 15 m and 215 m; the 5 m offset was due to a previous cable repair. The Jordan Basin bottom instrument, measuring temperature/conductivity as well as pressure, remained in place for the entire thirteen-month observation period.

The Northeast Channel mooring had a near-surface water temperature sensor, four temperature-conductivity sensor pairs located at 50 m intervals between 22 m and 162 m (Table 1). As with the Wilkinson Basin mooring, the sensor depth differences were due to earlier mooring cable repairs. This mooring was troubled with a pair of problems which fortunately were reported via telemetry. In response, the mooring was recovered in February, repaired and redeployed on 5 April 1987, functioning properly thereafter until its recovery in September 1987. The bottom instrument in the Northeast Channel was damaged—apparently due to a collision with a bottom-dragger—and not recovered.

3. Water property evolution

Seasonal changes in Gulf of Maine water properties are caused through exchanges with both the local atmosphere and the adjacent shelf/slope ocean. As revealed by the depth/time distributions of temperature and salinity isopleths in central Jordan Basin (Fig. 5), during autumn, wind-induced cooling significantly reduced vertical stratification in the western Gulf, setting the stage for wintertime deep convective mixing. The resulting deep mixed layer is evident in Wilkinson Basin water properties (Fig. 5). The entrainment of the deeper, saltier water associated with deep mixing left an imprint on mid-depth water column properties in the western Gulf for the rest of the year. During late autumn and early winter 1986 in the eastern Gulf, unusually large inflows of relatively fresh water from the Scotian Shelf helped to maintain a relatively stronger vertical stratification. Thus deep mixed layer formation was inhibited even more than usual in the eastern Gulf during winter 1986–87. The subsequent advection of the fresher eastern Gulf water into the western Gulf later in the winter helped to increase the stability and reduce vertical mixing there. Eventually, springtime 1987 surface warming and some local river runoff (like other years) increased the stability of the water column throughout the Gulf, thus isolating 1987 Maine Intermediate Water from the warmer Maine Surface Water. While the scenario described above repeats itself annually, it appears that the unusually large amount of freshwater introduced into the Gulf during winter 1986–87 was responsible for the anomalies seen in the evolution of water mass distribution patterns described herein and the density and geostrophic flow structure (Brown and Irish, 1992) during the spring/summer 1987.

a. Autumn/early winter: Preconditioning of the water column. As in other years, the summer 1986 water property and stratification structure was radically changed by the effects of autumn cooling. The time series of the vertical stratification strength (buoyancy frequency squared; N^2) averaged between about 15 m and 115 m in the

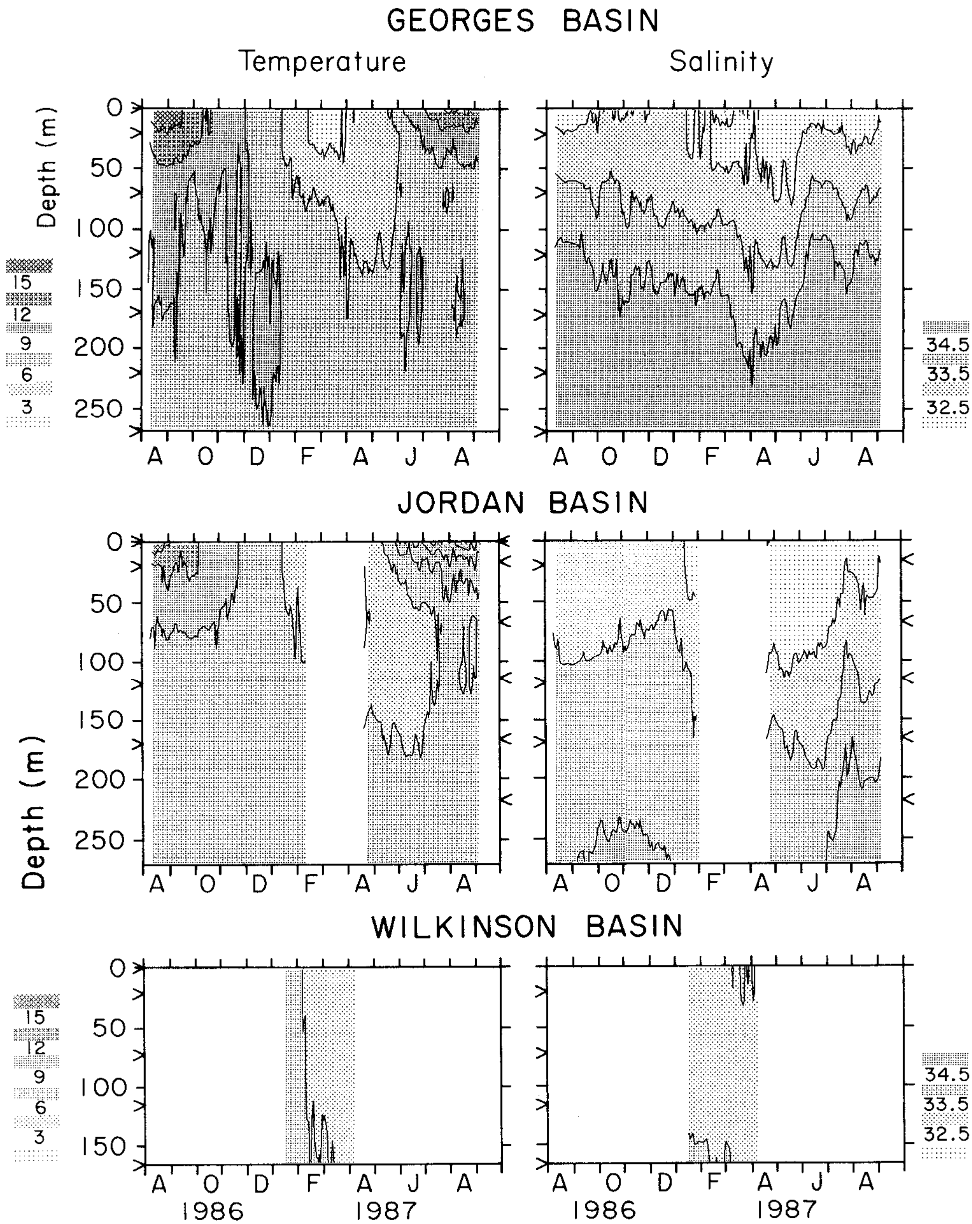


Figure 5. The evolution of temperature and salinity in Georges, Jordan, and Wilkinson Basins. These depth-versus-time contour plots of temperature (left) and salinity (right) were derived through linear interpolation between adjacent subtidal time series measurements at the depths of the carets. The legends for temperature ($\delta T = 3°C$) and for salinity ($\delta S = 1.0$ psu) are shown on the right and left respectively. The Wilkinson Basin observations started late and ended early when the mooring was moved to Jordan Basin.

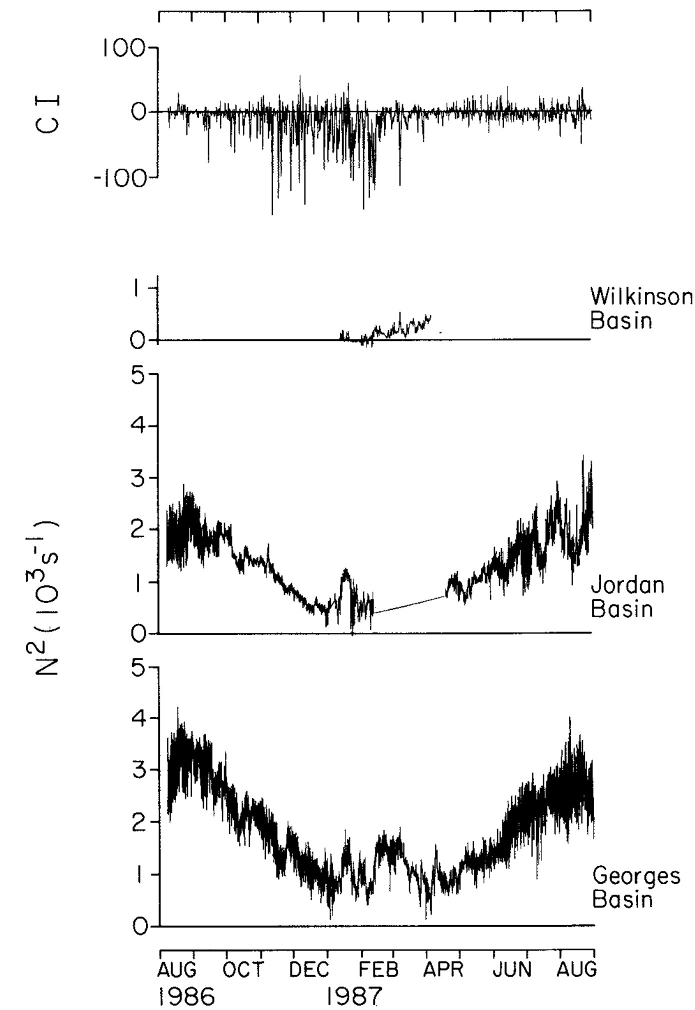


Figure 6. Stratification strength (N^2, buoyancy frequency-squared) time series in Georges, Jordan and Wilkinson Basins. N^2 values averages over the 15 m to 115 m depth interval. A cooling index (*CI,* (see text) time series is presented for comparison.

center of Georges and Jordan Basins exhibit a general decrease from their late August maxima to their respective minima in the January-February time frame (Fig. 6). We assume that, had they been available, autumn/early winter observations in Wilkinson Basin would have revealed a similar trend, though perhaps steeper than those observed.

The importance of cooling, in reducing the upper water column stratification strength, was assessed using a proxy for actual surface heat flux time series which we were unable to compute. The form of the proxy "cooling index" is based on the work of Brown and Beardsley (1978), who have shown that combined sensible and latent heat flux during periods of offshore cold winds are primarily responsible for extracting heat from the water column during late fall and early winter. Both of these heat flux components are generally estimated with bulk formulae which involve ocean/atmosphere temperature and water vapor pressure differences as well as wind speed. (We were not able to compute the heat flux time series due to a lack of water vapor pressure measurements.) Thus we define a cooling index based on the atmosphere minus ocean temperature difference δT and the offshore wind (toward 153°T) speed W_x according to

$$CI = \frac{W_x \delta T}{\sigma_{W_x} \sigma_{\delta T}},$$

where the normalization is the product of the respective standard deviations. (Thus negative CI corresponds to cooling of the ocean by cold offshore winds.) There is some justification in this approach because water vapor pressure differences depend strongly on temperature differences.

The Figure 6 comparison of cooling index and the 15 m–115 m water column stability (i.e., N^2) in both Jordan and Georges Basins clearly suggests the importance of heat extraction in the general erosion of water column stability during the autumn. (The results of Brown and Irish (1992) show that the underlying vertical stability induced by salinity stratification has N^2 values were about 1×10^{-4} s^{-2}.) A closer look at relatively rapid stability decreases suggests that a combination of mixing due to cooling and windstress eroded the water column stability. The series of intermittent vertical mixing events during autumn and early winter effectively reduced the vertical stratification strength and thus "preconditioned" the water column for the wintertime deep convection that was to follow.

The integrated effects of the autumn preconditioning and subsequent vertical mixing on stratification structure are illustrated in the Figure 7 comparisons of August 1986 and February 1987 buoyancy frequency squared (N^2) profiles. Note the extreme change in Wilkinson Basin, from the strongest stratification in August 1986 to the weakest in February 1987. The 130 m depth of the mixed layer (i.e., near zero N^2) is convincing evidence that deep mixing occurs preferentially in Wilkinson Basin. The mixed layer depth in central Jordan Basin was shallower (about 60 m) than in central Wilkinson Basin. Even though Jordan Basin surface waters were probably cooled as much or more than those in Wilkinson Basin, the added vertical stability associated with the relatively fresher water in the eastern Gulf undoubtedly inhibited the formation of a deeper mixed layer. The mixed-layer depths in Georges Basin were the shallowest (about 10 m) due to the combination of stronger vertical stratification and reduced heat extraction by the atmosphere already warmed after flowing over surface waters in the Jordan Basin region.

The autumn/early winter preconditioning of the eastern Gulf upper waters was

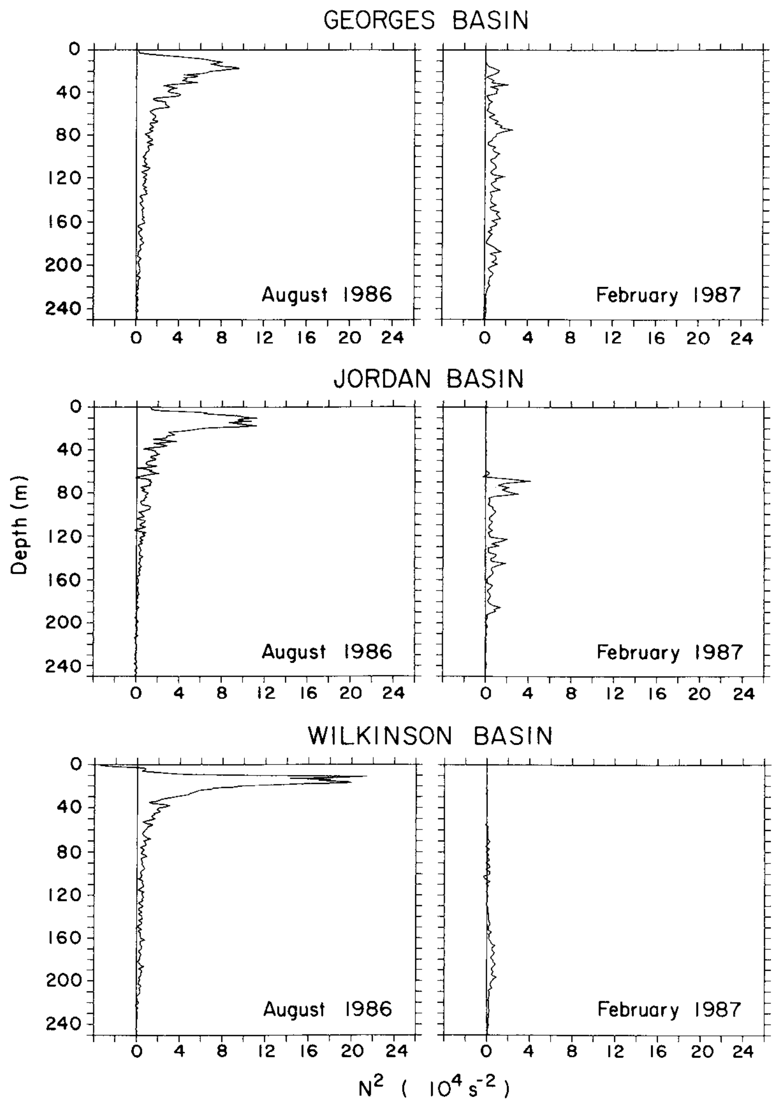


Figure 7. Stratification strength (N^2; buoyancy frequency squared) profiles in the three major basins in the Gulf for times of maximum (August 1986) and minimum (February 1987) strength. Each profile is an average of the CTD profiles obtained from each of these basins.

less effective than in the western Gulf due to the inflow of relatively fresh water from the Scotian shelf during late autumn and early winter. In November 1986, increased amounts of relatively fresh Scotian Shelf water began to enter the eastern Gulf south of Cape Sable (Smith, 1983) and through the Northeast Channel (Brown and Beardsley, 1978). Surface geostrophic flow estimates (Brown and Irish, 1992) and the salinity distribution histories (Fig. 5) support this conclusion. The salinity maps from

February 1987 (Fig. 8) clearly show influence of these relatively fresh inflows of Scotian Shelf Water on upper water column salinity in the eastern Gulf. The wintertime maxima in Georges and Jordan Basins N^2 time series (Fig. 6) were induced by the continued inflows of the fresher Scotian Shelf Water. Later, we will return to the issue of 1987 freshwater accumulation in the Gulf.

b. Winter water production. Winter water production increased considerably in early 1987. After the vertical stratification had been eradicated in the western Gulf, continued surface cooling produced the dense water that sank. Shallower sinking occurred in the coastal zone and deeper sinking occurred in the deeper basins. The observations described below are consistent with the suggestion of Mountain and Jessen (1986), namely, that under similar cooling conditions the somewhat fresher surface water in the region of Jordan Basin does not penetrate as deeply as does the saltier surface water in western Wilkinson Basin.

Convective overturning along coastal Maine in December and January has been shown (Brown and Beardsley, 1978) to entrain saltier deep water and produce elevated surface salinity patches inshore of Wilkinson and Jordan Basins. The Townsend *et al.* (1987) coastal CTD observations (integrated into Fig. 8) define just such a small-scale saltier (and "warmer") patch just east of Portland in February 1987. We surmise that this feature was newly upwelled deeper water associated with the cooling-induced sinking of the fresher water patch located to the southeast. The subsurface signature of this colder, fresher water downdraft is seen in the 100 db salinity section (Fig. 8). Indeed, the property distribution of this region suggests significant sinking and spreading of relatively cold, fresh water around the northwestern edge of Jordan Basin.

The effects of deeper convection are evident in central Wilkinson Basin observations in February 1987. The intermittent negative vertical stratification values over the upper 100 m in late January (Fig. 6) document deep convection events which reoccur there through mid-February. Water property distribution sections (Fig. 9) show that deep convective mixing in 1987 extended to depths approaching 200 m in western Wilkinson Basin. The location of the thickest mixed layer suggests that the central basin overturning process in 1987 appears to have been augmented by downslope flow of anomalously dense water formed perhaps in Massachusetts Bay. Mountain and Jessen (1986) suggest that the deep mixing occurs only during some winters, when the combination of surface cooling and salinity stratification are favorable. 1987 appears to have been such a year.

Observations show that convective overturning in Wilkinson Basin was important to the formation of large amounts of Winter Water during 1987. The January-to-April range of T-S properties in central Wilkinson Basin is defined by the envelope of subtidal time series T-S observations (dashed line in Fig. 10). The comparison of a pair of T-S relations (Fig. 11) for the water column at the Wilkinson Basin mooring

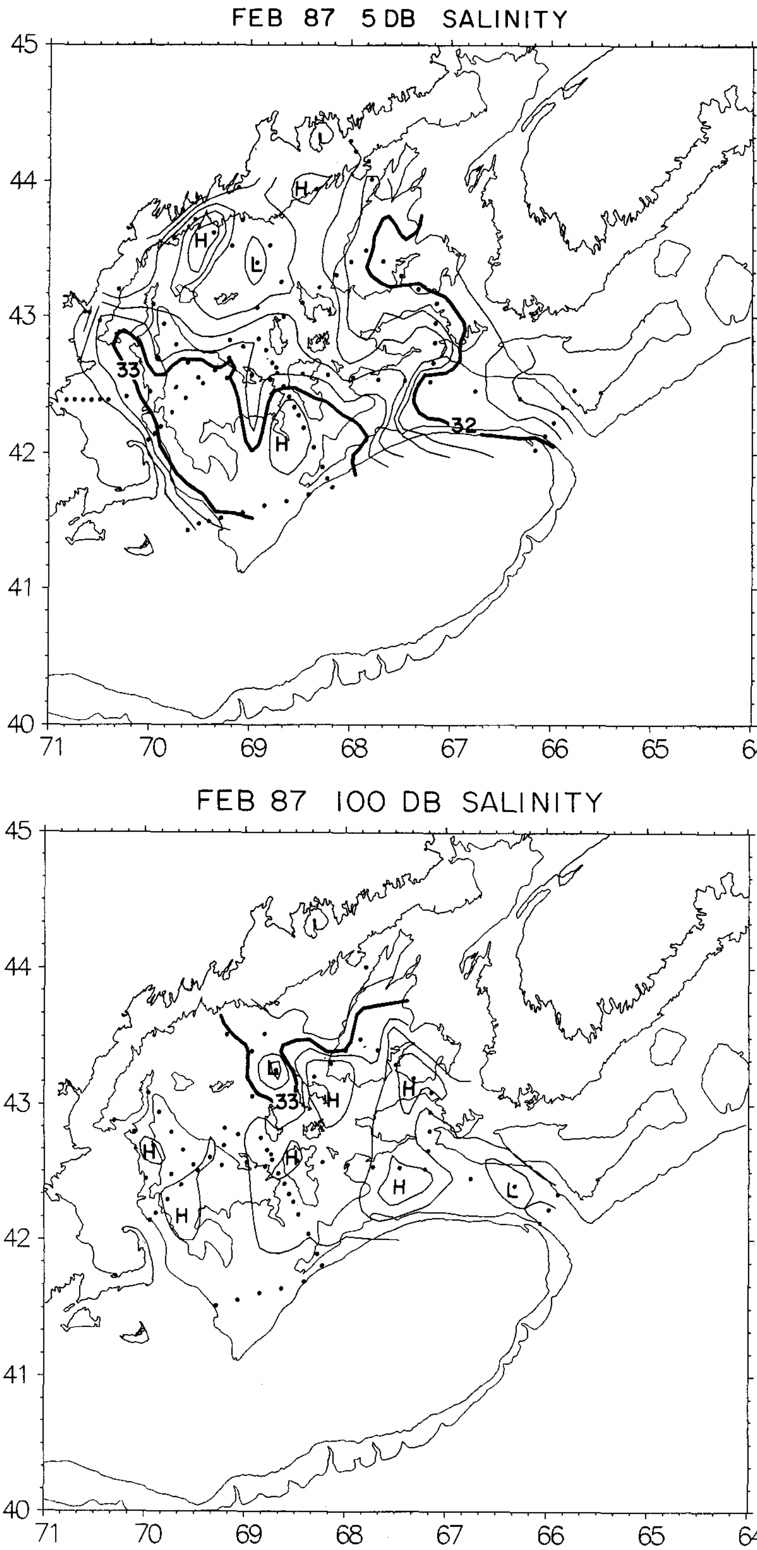


Figure 8. Horizontal distributions of 5 db and 100 db salinity in the Gulf of Maine for the period 5–20 February 1987. The contour interval is 0.2 psu with local salinity maximas with minimas indicated.

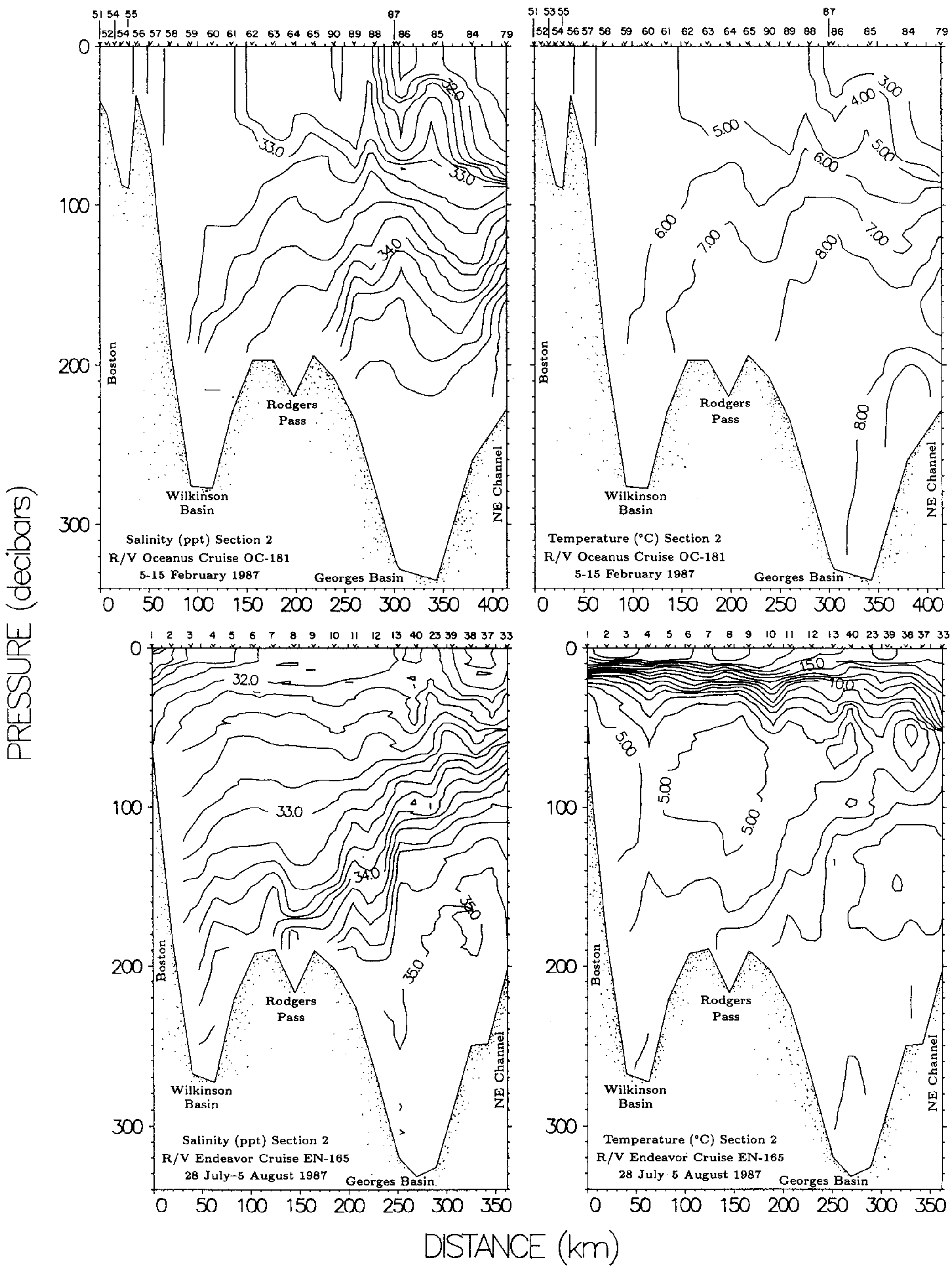


Figure 9. A comparison of temperature (right) and salinity (left) sections between Wilkinson Basin to Northeast Channel from winter (above) and summer (below). Contour intervals are 1°C for temperature and 0.2 psu for salinity.

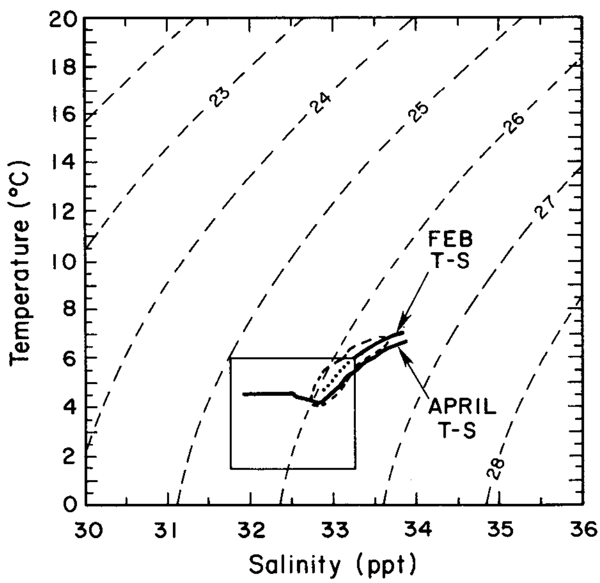


Figure 10. A comparison of temperature-salinity (T-S) relationships derived from shipboard and moored observations in Wilkinson Basin during winter and early spring 1987. The dashed line envelopes the subtidal moored T-S time series. CTD-derived T-S relations for 7 February (solid), 15 February (solid plus dotted), and April near the Wilkinson Basin mooring site are compared. The box defines the 1987 Maine Intermediate Water T-S characteristics (see text).

site on 7 and 15 February, respectively, show increased amounts of cooler (and fresher!) water at depth. We conclude that an intrusion of colder, fresher and actually less dense near-surface water in early February was responsible for expanding the time series T-S envelope toward the lower left. That intrusion was part of the Gulf response to a severe winter "northeasterly" on 11 February 1987. The water mass, which moved into central Wilkinson Basin during this storm was stable enough to inhibit any further deep convective mixing (i.e., no negative N^2 values) in *central* Wilkinson Basin for the rest of the 1987 cooling season (see Fig. 6). Nevertheless, continued cooling of the near-surface water column in and around the basin plus advection of the cooled water into central Wilkinson Basin between February and April led to the observed downward expansion of the T-S property envelope there. During March, the seasonal minimum water column temperatures were reached throughout the Gulf, signaling the end of winter water formation and the establishment of mid-depth water properties for 1987. Before proceeding with the spring and summer 1987 water mass evolution story, however, we return to the issue of the freshwater contributions to the Gulf during 1987.

c. Freshwater input and distribution: 1986–87. The two principal sources of Gulf freshwater are the wintertime inflows of the relatively fresh Scotian Shelf Water and

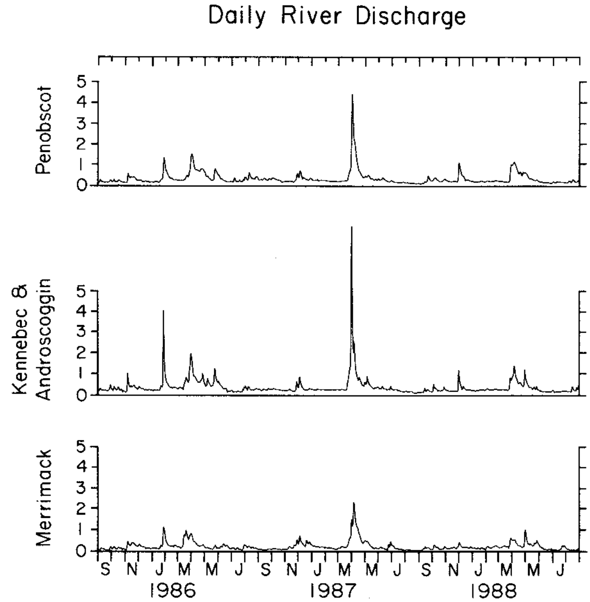


Figure 11. River discharge rates (100 $m^3 s^{-1}$) measured at USGS gauging stations on the four largest U.S. rivers emptying directly into the Gulf of Maine. The Kennebec and Androscoggin Rivers empty into the same estuary and thus are combined.

the combined discharge of rivers directly into the Gulf. According to data from Budyko (1963) and Bunker (1976), the precipitation (P) minus evaporation (E) over the Gulf of Maine is about 20 cm/year. Thus the annual rate of P-E freshwater addition is only about 25% of the combined local rivers and of course is distributed differently. Freshwater is exported from the Gulf along with Maine Surface and Intermediate Water. Given the different inflow/outflow senarios, the amount of freshwater can accumulate or diminish seasonally.

The changes in the distributions and volumes of freshwater within the Gulf are described in terms of the freshwater (presumably of "coastal" origin) required to dilute ocean water with a salinity S_o. The reference salinity of ocean water in this case is 35.25 psu, the maximum salinity observed in the study. A useful variable in this regard is the freshwater index $F = (S_o - S)/S_o$ which varies from 1 to 0 as the observed salinity(s) varies from 0 to S_o. Vertical integrals of F over the water column yield equivalent thicknesses of "pure freshwater." Horizontal distributions of freshwater thickness for the full water column were integrated laterally to obtain total freshwater volumes for each of the CTD surveys (Table 2). To account for the differing coverage areas in each survey when comparing results, the total freshwater volumes of the Gulf for each survey were normalized according to the April 1987 survey volumes. The freshwater volume histories have been determined for the full survey volume as well as for just the upper 40 m. The 0–40 m freshwater volumes are

Table 2. Freshwater volumes in both the upper 40 m and the full water column of the Gulf of Maine for each of the 1986–87 CTD surveys. For comparison purposes, the volumes (10^{10} m^3) have been normalized to the corresponding volume surveyed in April 1987. Volume changes and time-interval-averaged freshwater transports T_F (10^3 m^3 s^{-1}) required to produce the volume changes are given for selected time intervals. The asterisk identifies the 1986–87 annual results derived by interpolation between observed results.

Ship survey date	Days	0–40 m Vol.	0–40 m Vol. change	0–40 m T_F	Gulf Vol.	Gulf Vol. change	Gulf T_F
12 Aug '86		24.98			71.80		
	184		2.27	1.43		5.60	3.52
12 Feb '87		27.24			77.39		
	59		2.41	4.73		16.00	31.39
12 Apr '87		29.65			93.40		
	111		1.19	1.24		−7.26	−7.57
1 Aug '87		30.84			86.13		
	42		−1.82	−5.02		−3.72	−10.25
12 Sept '87		29.02			82.42		
12 Aug '86–12 Apr '87	243		4.68	2.23		21.60	10.28
12 Aug '86–12 Sep '87	396		4.05	1.18		10.62	3.10
12 Aug '86–12 Aug '87*	365		5.36*	1.70*		13.29*	4.21*

more sensitive to the contributions from the local Gulf of Maine rivers than are the full water column freshwater volumes.

The comparison of freshwater volumes clearly (Table 2) shows net accumulation of freshwater within the Gulf between August 1986 and 1987. The greatest increase in total freshwater volume within the Gulf occurred between February and April 1987. The net loss of freshwater from the Gulf during the spring and summer was not enough to return the previous year's lower levels of freshness. The upper 40 m freshwater senario was somewhat different from that for the Gulf as a whole. While the greatest increase occurred between February 1986 and April 1987, the volume continued to increase until August. The local river inputs during April and May probably delayed the decrease in 0–40 m freshwater until after August.

The local freshwater comes from Canadian and U.S. rivers which discharge about equal amounts of freshwater directly into the Gulf (Emery and Uchupi, 1972). To determine the direct river runoff to the Gulf in 1987, we obtained daily USGS discharge data for the four major U.S. rivers (70% of U.S. discharge) and monthly discharge data for the St. John River (65% of Canadian discharge). Despite the prominent April 1987 peaks in the U.S. river discharge (Fig. 11), the August 1986–87 average combined U.S. river discharge (1064 m^3 s^{-1}) was about the same as the

Table 3. Average combined discharge rates (10^3 m^3 s^{-1}) of the Penobscot, Kennebeck, Androscoggin and Merrimack Rivers (PKAM) for comparable time periods during 1985–86 and 1986–87 respectively. The rates of total 'local' river discharge into the Gulf (T) were estimated by scaling up PKAM by 2.72; based on Emery and Uchupi (1972).

		1985–86		1986–87	
Interval	Days	PKAM	T	PKAM	T
12 Aug–12 Feb	184	0.81	2.20	0.77	2.09
12 Feb–12 Mar	28	0.85	2.31	0.66	1.79
12 Mar–12 Apr	31	2.28	6.20	3.65	9.92
12 Feb–12 Apr	59	1.60	4.35	2.23	6.06
12 Apr–1 Aug	111	1.02	2.77	1.01	2.74
1 Aug–12 Sep	42	0.79	2.15	0.31	0.84
12 Aug–12 Sep	396	0.99	2.69	1.00	2.72
12 Aug–12 Aug	365	1.01	2.75	1.06	2.88

August 1985–86 average river discharge (1007 m^3 s^{-1}). In fact, the average discharge rates from the U.S. rivers in both 1985–86 and 1986–87 were slightly less than the long-term average discharge rate of 1109 m^3 s^{-1} (Emery and Uchupi, 1972, using data from Bue, 1970, and Wilson and Iseri, 1969). With the exception of the 12 March–12 April period, the average discharge rates for a selected set of 1987 periods were actually less than the average discharge rates for corresponding periods in 1985–86 (Table 3). Like the U.S. rivers, the Canadian rivers discharged somewhat less freshwater than normal in 1987. Thus we conclude that the anomalous accumulation of freshwater in the 1986–1987 Gulf was due to inflow of Scotian Shelf Water.

The largest amounts of freshwater entered the Gulf in late January and early February around Cape Sable *and* through the Northeast Channel (Figs. 8, 12). This surge of freshwater inflow was very likely buoyancy-driven and due to the largest September through January St. Lawrence River discharge of the century (Mountain, 1991). It may be anomalous events like this that also move Scotian Shelf Water directly onto Georges Bank. The inflow is reflected clearly in the large February to April 1987 increase in the total freshwater volume (Table 2). The August 1986 through April 1987 sequence of 0–40 m maps (Fig. 12) suggests that it took some time for the fresher portion of the Scotian Shelf Water to arrive in Jordan Basin. It is unlikely that the peak local river discharge during the late March/early April interval (Fig. 11) contributed in a significant way to the central Gulf freshwater pool in mid-April.

Despite the dominant contribution of Scotian Shelf inflow to the overall freshwater budget of the Gulf, local rivers do make significant contributions to the freshwater budget of the upper 40 m. This conclusion is suggested by a comparison of the freshwater "demand" rates of the Gulf for different time periods (Table 2) and the potential river discharge "supply" rates (Table 3). Between April and August 1987, the 0–40 m freshwater volume increased, while the overall Gulf freshwater volume

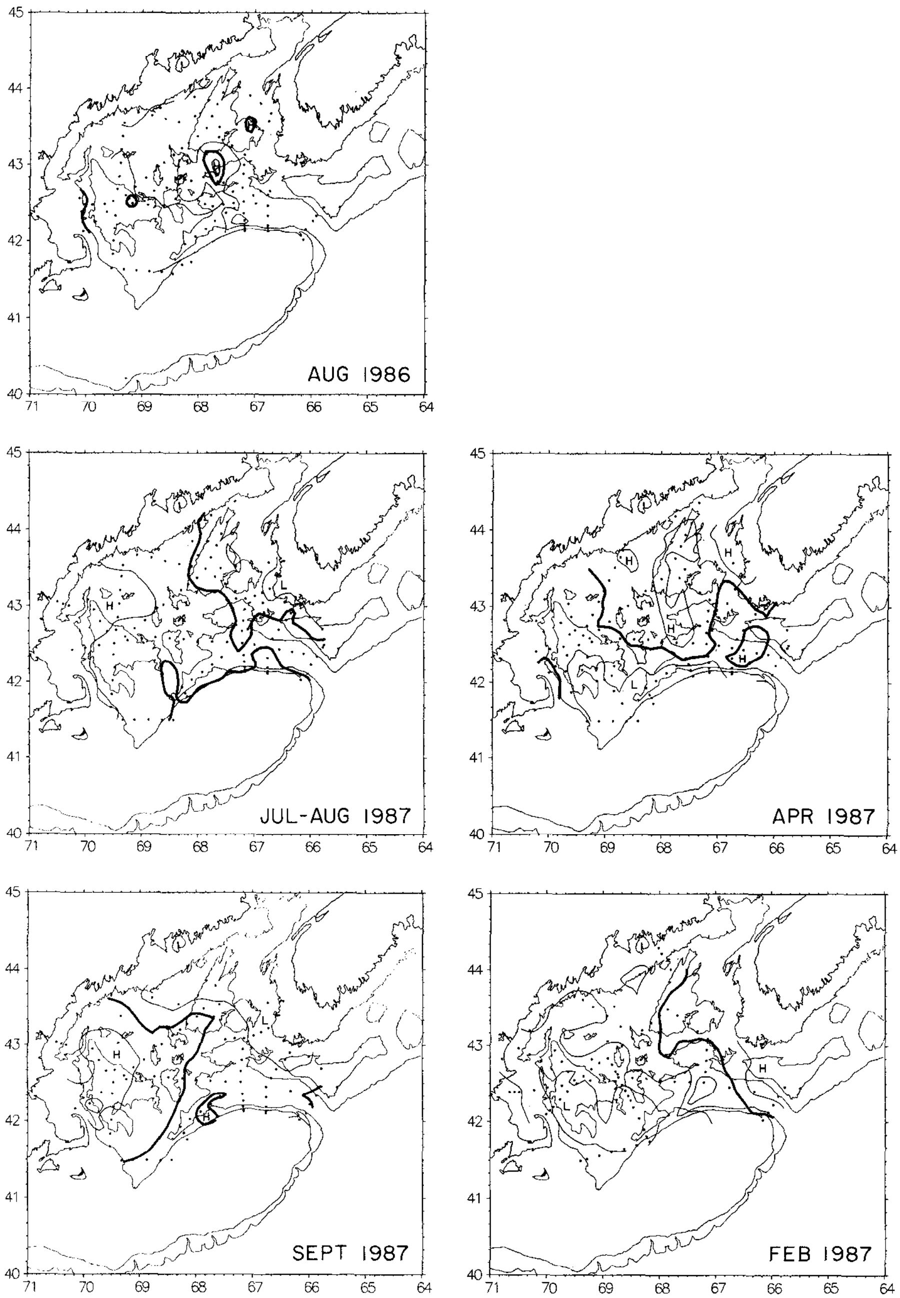


Figure 12. An August 1986 through September 1987 sequence of 0–40 m fresh-water thickness distribution maps. The contour interval is 0.5 m with a highlighted 3.5 m thickness contour. The maximas and minimas are indicated.

decreased. The volumetrics (Tables 2 and 3) and the July–August map (Fig. 12) strongly suggest that the March/April river discharge pulse contributed to that summertime increase of upper ocean freshwater, most of which was in the western Gulf in early August. It is worth noting that the 1987 local river discharge was essentially normal for this period. The general export of water from the western Gulf through the Great South and/or Northeast Channels undoubtedly was responsible for the August to September decrease in the total Gulf freshwater volume. But, as we will see, the excess of freshwater in the 1987 Gulf had direct and indirect influences on the water mass evolution story which we develop next.

4. Water mass evolution

The 1987 Gulf of Maine water masses have been defined in terms of Maine Intermediate Water—the water mass whose cold, fresh characteristics are particularly sensitive to the details of the inflow and winter water production during the previous winter. Maine Surface Water and Maine Bottom Water were then defined assuming that no mixtures of these water masses and Maine Intermediate Water existed in April 1987. Maine Bottom Water and Slope Water definitions were based on less precise historical definitions and our own observations.

The definition of Maine Intermediate Water ('Intermediate Water', henceforth) for a particular year is very sensitive to the amount of winter cooling and to the timing and volume of the preceding winter's inflows and outflows, as discussed above. Following Hopkins and Garfield (1979), we have defined 1987 Intermediate Water properties in terms of the statistics of the April 1987 water column temperature minimum $\overline{T}_{min}$ according to $\overline{T}_{min} \pm 2\sigma_T$ and $\overline{S}_{min} \pm 2\sigma_S$, where $\overline{T}_{min}$ is the Gulf-wide average of the water column temperature minimums between 40 m and 150 m depth, $\overline{S}_{min}$ the average of the associated salinities, and the σ's are the corresponding standard deviations. The 1987 Intermediate Water property envelope (Table 4) compares favorably with the Hopkins and Garfield (1979) Intermediate Water envelopes for 1965, 1966, and 1970. The fact that 1987 Intermediate Water covers a somewhat wider range of property values than the 1965 and 1966 Intermediate Water may be related to (a) the greater extent and better spatial resolution of our 1987 stations, and/or (b) the fact that the observations used for the computation were from April rather than May like the others. The disadvantage with using April observations to define 1987 Intermediate Water is that surface warming had not yet produced a distinct Surface Water mass (see April *T-S* relation in Fig. 10). Fortunately, as the 1986–87 time series observations (Fig. 4) show, minimum temperatures at depths of 70 m had been reached in each of the basins during March 1987. Because Winter Water was no longer being formed, the Hopkins and Garfield (1979) approach represents a reasonable *guide* for defining Intermediate Water.

Table 4. The Hopkins and Garfield (1979) Maine Intermediate Water definitions for several different years. Maine Intermediate Water is defined in terms of the temperature range $\overline{T}_{min} \pm 2\sigma$ (where T_{min} is the minimum temperature between 40 m and 150 m depth and σ is the standard deviation of all T_{min}) and the salinity range $\overline{S}_{min} \pm 2\sigma$, where S_{min} is the salinity corresponding to the minimum temperature. Note that the considerably smaller 1976 volume pertains only to the basins and does not include surrounding regions as do the others.

	Mean		Range		Volume	
Month/year	*T*	*S*	*T*	*S*	km^3	Authors
May 1965	3.20	32.44	2.20–4.20	31.97–32.91	5800	H&G (1979)
May 1966	3.96	32.67	3.10–4.90	32.01–33.23	5500	H&G (1979)
May 1970	4.99	32.57	3.60–6.30	31.53–33.61	6500	H&G (1979)
May 1976	6.00	32.90	4.50–7.50	32.20–33.60	2687	Schlitz *et al.* (1984)
April 1987	3.61	32.47	2.01–5.21	31.73–33.21	6420	This study

a. Water mass definitions: 1987. Maine Intermediate Water for 1987 was defined by a Hopkins/Garfield-derived salinity range and a slightly modifed temperature range. Specifically, we adopted the 31.75 psu to 33.25 psu salinity range indicated on the composite of *T*/*S* relationships from April 1987 CTD observations within the Gulf of Maine proper (Fig. 13). (Northeast Channel water *T*/*S* relations were excluded from the composites because they were clearly influenced by water that had not and might never have entered the Gulf at the time of the April survey.) To ensure that all water in the Gulf in April was part of a defined water mass, the 1987 Intermediate Water temperature range was expanded from the Hopkins/Garfield-derived range of 2.01–5.21°C (Table 4) to 1.5°C–6.0°C (see April, Fig. 13). This approach enabled us to account for subsequent water mass volume changes and thus estimate the mixing rate between water masses during the spring and summer, as discussed later.

Strictly speaking, Maine Surface Water ('Surface Water', henceforth) during April (and some of May) 1987 consisted only of water that was fresher than 31.75 psu. Thus, while most of the near-surface water in the Gulf during April had the properties of Intermediate Water, it was gradually converted to Surface Water during May. The upper salinity limit of 32.5 psu for Surface Water warmer than 6°C was chosen because that isohaline was found to define the approximate depth of the base of the summer 1987 thermocline and hence pycnocline. Thus the 32.5 psu isohaline marked the summertime separation between Surface Water and Intermediate Water or its mid-depth replacement water mass.

The definition of Maine Bottom Water ('Bottom Water', henceforth) adopted here reflects the fact that it is derived from mixtures of deep Slope Water (see below) and mid-depth water masses. The lower salinity boundary of 33.25 psu for 1987 Bottom Water was set by the Intermediate Water definition (see above). The Bottom Water/Slope Water salinity boundary of 34.5 psu was based on the lower salinity limit on Laborador Slope Water indicated by Hopkins and Garfield (1979). This

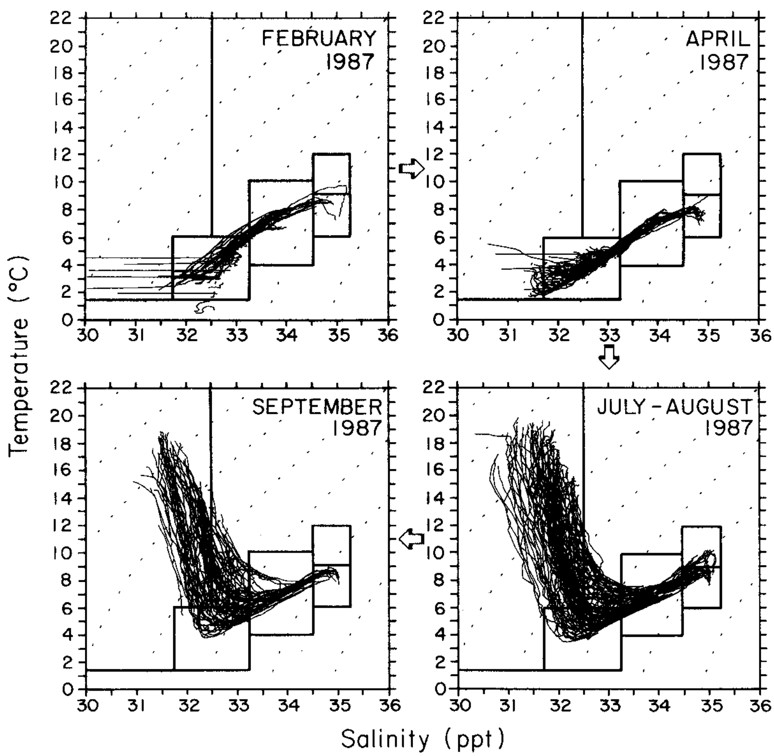


Figure 13. A comparison of temperature-salinity (T-S) relation composites for each of the CTD surveys of the Gulf of Maine between February and September 1987. The T-S relations for Northeast Channel CTD casts have been omitted. The rectangles superposed on each composite define the T-S properties of 1987 Maine Intermediate Water (*MIW*), Maine Surface Water (*MSW*), Maine Bottom Water (*MBW*), Warm Slope Water (*WSW*), and Labrador Slope Water (*LSW*) as described in the text.

choice is also reasonably consistent with Slope Water salinity ranges used by Ramp *et al.* (1985) and Gatien (1975). An upper temperature limit of 10°C was adopted for 1987 Bottom Water. This limit is reasonably consistent with historical definitions and further it meant that all Bottom Water with salinities in the defined salinity range (33.25 psu $< S_{MBW} <$ 34.5 psu) would be classified as "pure" Maine Bottom Water in both August 1986 and September 1987. We will return to this point later. The lower temperature limit of 4°C for 1987 Bottom Water is somewhat arbitrary and in fact could have been set at 4.5°C with no effect on the following analysis.

The definitions of the two kinds of Slope Water are based on historical Slope Water definitions as well as with the practical consideration of accounting for all Gulf of Maine deep water in terms of Bottom and Slope Water. The 34.50 psu lower salinity limit separates Slope and Bottom Water (see above). The 35.25 psu upper salinity limit is somewhat crude but accounts for all 1987 Slope Water. The 9°C

Table 5. Gulf of Maine water mass definitions for 1987. Maine Surface Water (*MSW*), Maine Intermediate Water (*MIW*), Maine Bottom Water (*MBW*), Warm Slope Water (*WSW*), and Labrador Slope Water (*LSW*) are defined on the basis of water property distributions in April 1987, as discussed in main text. The depth criterion was applied in determining April MSW and MIW volumes only.

Water	Salinity (ppt)		Temperature (°C)		Depth (m)		
masses	Max	Min	Max	Min	Max	Min	
MSW	32.50	28.00	25.0	6.0			
	31.75	28.00	6.0	1.5			
	33.25	31.75	6.0	1.5	40	0	April–May only
MIW	33.25	31.75	6.0	1.5	—	40	April–May only
MBW	34.50	33.25	10.0	4.0			
WSW	35.25	34.50	12.0	9.0			
LSW	35.25	34.50	9.0	6.0			

temperature division between cold Labrador Slope Water and Warm Slope Water is based on Gatien's (1975) work. The 12°C upper temperature limit for Warm Slope Water and 6°C lower temperature limit for Labrador Slope Water are loosely based on historical definitions, while enabling us to account for all of the 1987 Slope Water.

The 1987 Gulf of Maine water mass definition framework based on the April observations (Table 5) has been superposed on the *T-S* relation composites from April as well as the other CTD surveys (Fig. 13). A comparison of the February and April composites reflects the general cooling and freshening of the Gulf during this time interval, as discussed in Section 3. The effects of spring and early summer warming of surface water and the inflow of Warm Slope Water are clearly seen in the July-August *T-S* composite. Note that Warm Slope Water is not present in the September *T-S* composite which excludes observations from the Northeast Channel as discussed above.

The summer *T-S* composites differ from the April composite in that they indicate the presence of an undefined intermediate depth water mass which fills a *T-S* niche partially surrounded by the defined water masses. Inspection of the individual *T-S* curves indicates that the warmer, saltier intermediate water is located in the eastern Gulf. Hereafter, this undefined water mass will be referred to as Summer Intermediate Water, consistent with its time of appearance. For purposes of the discussion to follow, we define 1987 Summer Intermediate Water to be all water with 32.5 psu $< S <$ 34.5 psu and 6°C $< T <$ 25°C.

Like other years many of the 1987 Maine water mass definitions are clearly related to their 1987 formation histories. Thus it is not surprising that a comparison of the August 1986 (Fig. 14) and the July/August 1987 (Fig. 13) *T/S* composites clearly shows that the 1987 Gulf of Maine was colder and fresher than the 1986 Gulf. It follows that the 1986 definitions of Surface, Maine Intermediate, and Bottom Water would be somewhat different from the corresponding 1987 definitions.

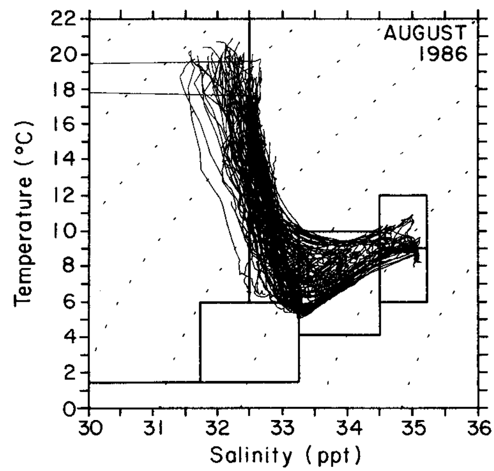


Figure 14. A composite of the August 1986 temperature-salinity (T-S) relations. The *T-S* relations for Northeast Channel CTD casts have been omitted. The rectangles superposed on the composite define the *T-S* properties of 1987 water masses as in Figure 13.

b. Water mass distributions. The moored temperature/salinity observations in the centers of Jordan and Georges Basins and in the Northeast Channel provide detailed temporal information concerning both water property and water mass structure during the time intervals between the CTD surveys. Month-long segments of subtidal temperature-salinity time series from Jordan Basin are presented in a sequence of *T-S* diagrams (Fig. 15). The May and June time series *T-S* relations document how surface warming creates the thermocline/pycnocline the separation between 1987 Maine Surface and Intermediate Water masses during the time period between the CTD surveys.

These time series *T-S* presentations differ from the instantaneous CTD-derived *T-S* relationships (e.g., Fig. 13) in that they show the range of *T-S* properties during the selected month at *only* the depths of the moored observations. There actually is some resemblence between CTD-derived and time series *T-S* relationships, when (a) the vertical structure of water properties is reasonably constant over the particular one-month period *and* (b) there is enough internal wave activity to sweep most of the local water column vertically past the set of sensors. This set of circumstances is apparently revealed in the composite April time-series *T-S* relations for Jordan Basin (Fig. 15) which compares well with an April CTD-derived *T-S* relationship from the mooring site. By contrast, the July time series *T/S* relations (e.g., 65 m), as we will show later, are due to the lateral inflow of nearly isopycnal water with the properties of water (e.g., Summer Intermediate Water) from outside our observation domain.

The water mass structure histories were constructed through the linear interpolation of observed *T/S* properties at the mooring sites (Fig. 16). These pictures clearly detail the timing of water mass structural change at these key observational sites. The

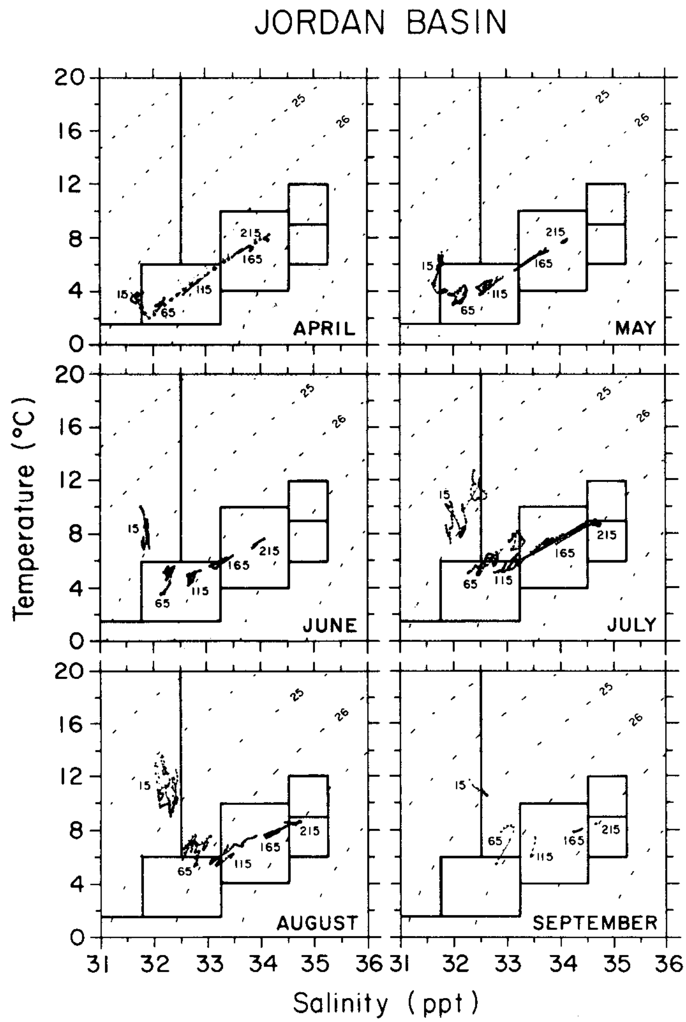


Figure 15. A sequence of monthly time-series temperature-salinity (*T-S*) relations in central Jordan Basin during 1987. One-month segments of the subtidal *T-S* time series at the indicated depths are presented in separate T-S diagrams. A CTD-derived *T-S* relation for April (dotted) is superposed on the April diagram. The 1987 water mass definition framework (see text) is also presented.

Northeast Channel history is critical in documenting the timing of inflows and outflows of both types of Slope Water and both types of Intermediate Water. There is evidence of an April/May outflow of Maine Intermediate Water through the Northeast Channel followed by further outflow during July and August (Fig. 16). In the interim there are clear suggestions of Slope Water inflow through the Northeast Channel, accompanied by some Summer Intermediate Water (see below). Inflow

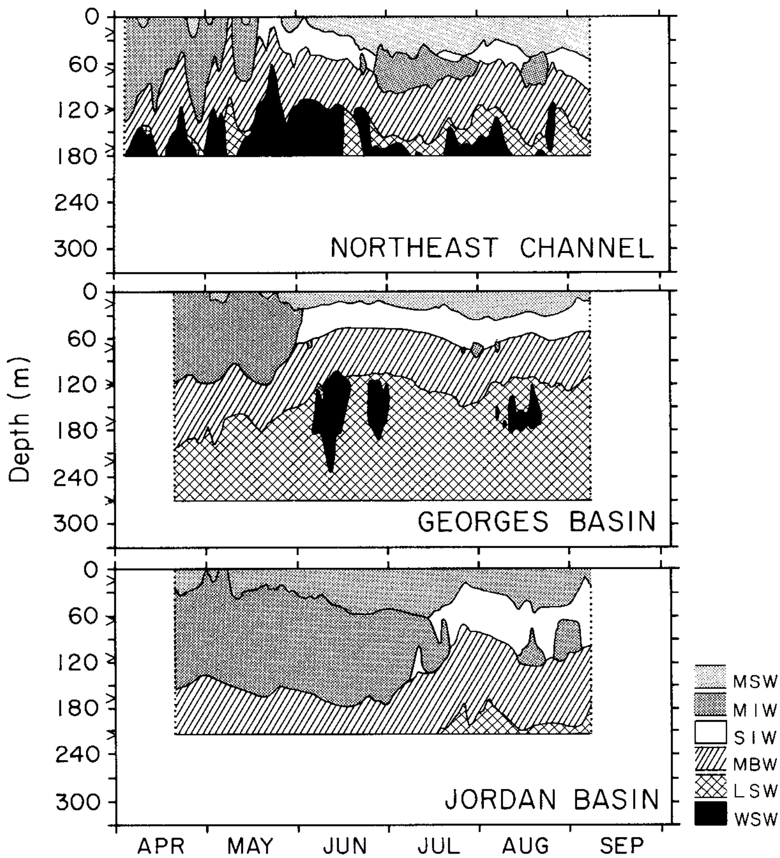


Figure 16. Water mass vertical structure time series from the Northeast Channel, Georges Basin and Jordan Basin mooring sites. The water mass structure (see legend) was determined by linear interpolation between observed temperature and salinity time series at the depths indicated by the carets.

pathways of Slope Water can be more clearly inferred by comparing water mass structure time series from the Northeast Channel and Georges Basin (Fig. 16). Warm Slope Water was the dominant deep water mass at the 170 m sensor depth in the Northeast Channel between the April through June. Labrador Slope Water became more dominant at 170 m as summer progressed. A comparison of time series water mass information suggests that a significant parcel of Warm Slope Water flowed into the Gulf during spring and eventually made its way to central Georges Basin by June.

The water mass structure history in the Northeast Channel also suggests that Summer Intermediate Water, initially at the surface and later at mid-depth, appears to have entered the Gulf from the Scotian Shelf/Slope region. The replacement of Maine Intermediate Water by Summer Intermediate Water during late May in

Georges Basin and July in Jordan Basin accompanied the sequential increases in the local thickness of Slope Water at each of the sites, suggesting a wholesale inflow. An analysis described later shows that Summer Intermediate and Slope Water inflows occurred at the same time as Maine Intermediate Water outflow from elsewhere in the Gulf.

More detailed quasi-synoptic pictures of the three-dimensional water mass structure were obtained for the April, July–August and September 1987 CTD surveys (Figs. 17–19). One of the more conspicuous features of the April vertical sections (Fig. 17) is the relatively large amounts of Maine Intermediate Water which reached the surface throughout most of the Gulf and the depth of 200 m in some parts of Wilkinson Basin. As documented by the moored time series (Fig. 16), the near-surface layer of Maine Intermediate Water was converted to Surface Water via warming during May. The subsequent replacement of Maine Intermediate Water by Summer Intermediate Water and the obvious influence of Slope Water inflows and movement are clearly depicted in these "snapshots" (Figs. 18 and 19). The transient nature of Warm Slope Water is also clearly seen in the seasonal sequence of transects.

The temporal and spatial coverage of the CTD surveys enabled us to document the 1987 changes in the volumes of the Gulf of Maine water masses. The actual measured volume of Maine Intermediate Water in April was a misleading overestimate of its "initial" 1987 volume because April near-surface Maine Intermediate Water was soon to be converted into Surface Water. Therefore, we reclassified all April Intermediate Water in the upper 40 m as Surface Water (see Fig. 17). The July–August water mass sections (Fig. 19) show that the choice of 40 m depth was a reasonable approximation for the late May Surface/Intermediate Water boundary.

The spring/summer 1987 evolutions in the distribution of the principal water masses in the Gulf are documented in terms of a set of contour maps of water mass thickness (Figs. 20–25). In the following section, we use these maps, the time series and other shipboard survey results to document and model the volume changes in the different water masses in the Gulf of Maine during 1987.

c. *Water mass volume changes.* The volumes of the different water masses have been estimated for the latter three surveys in 1987 by summing the gridded thickness values used to make the water mass distribution maps (Figs. 20–25). The total volumes of the water surveyed during each of the cruises varied by about 20% (see Table 6) because of the somewhat different area covered by each survey. To facilitate comparisons between survey results, *all* water mass volumes were normalized to yield the 13.09×10^{12} m^3 total volume of the April survey. This volume represents 81.5% of the total volume of the Gulf as determined by adding the Schlitz *et al.* (1984) volume estimate for the Gulf of Maine proper to our 0.6×10^{12} m^3 estimate for the Bay of Fundy.

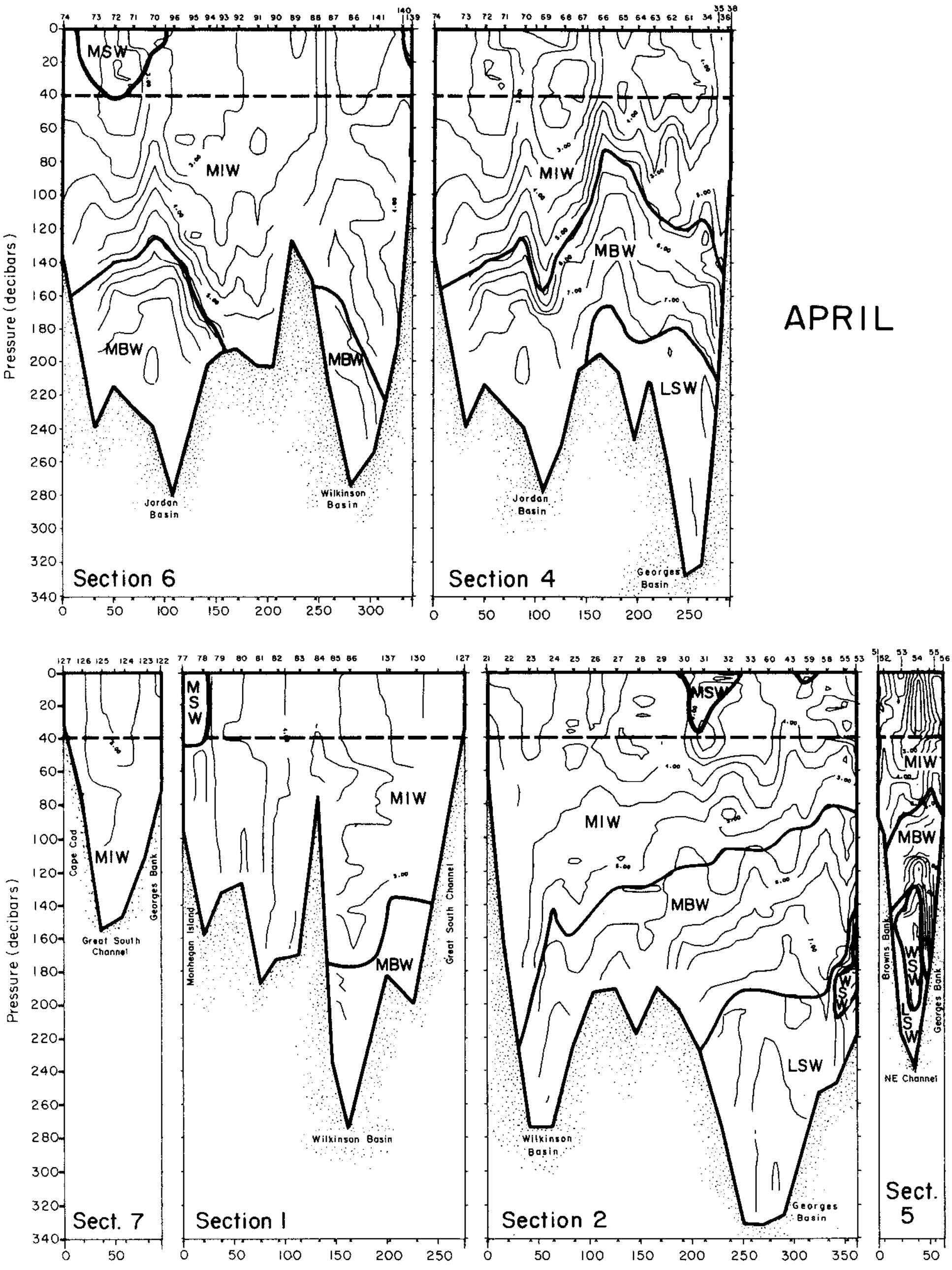


Figure 17. Water mass and temperature distribution sections for April 1987. Sections 1, 2, 4, and 6 crisscross the Gulf (see Fig. 2), while sections 5 and 7 are located in the Northeast and Great South Channels respectively. The dashed lines indicate the artificial boundary used to estimate the initial volumes of Maine Surface Water (MSW) and Maine Intermediate Water (MIW) (see text).

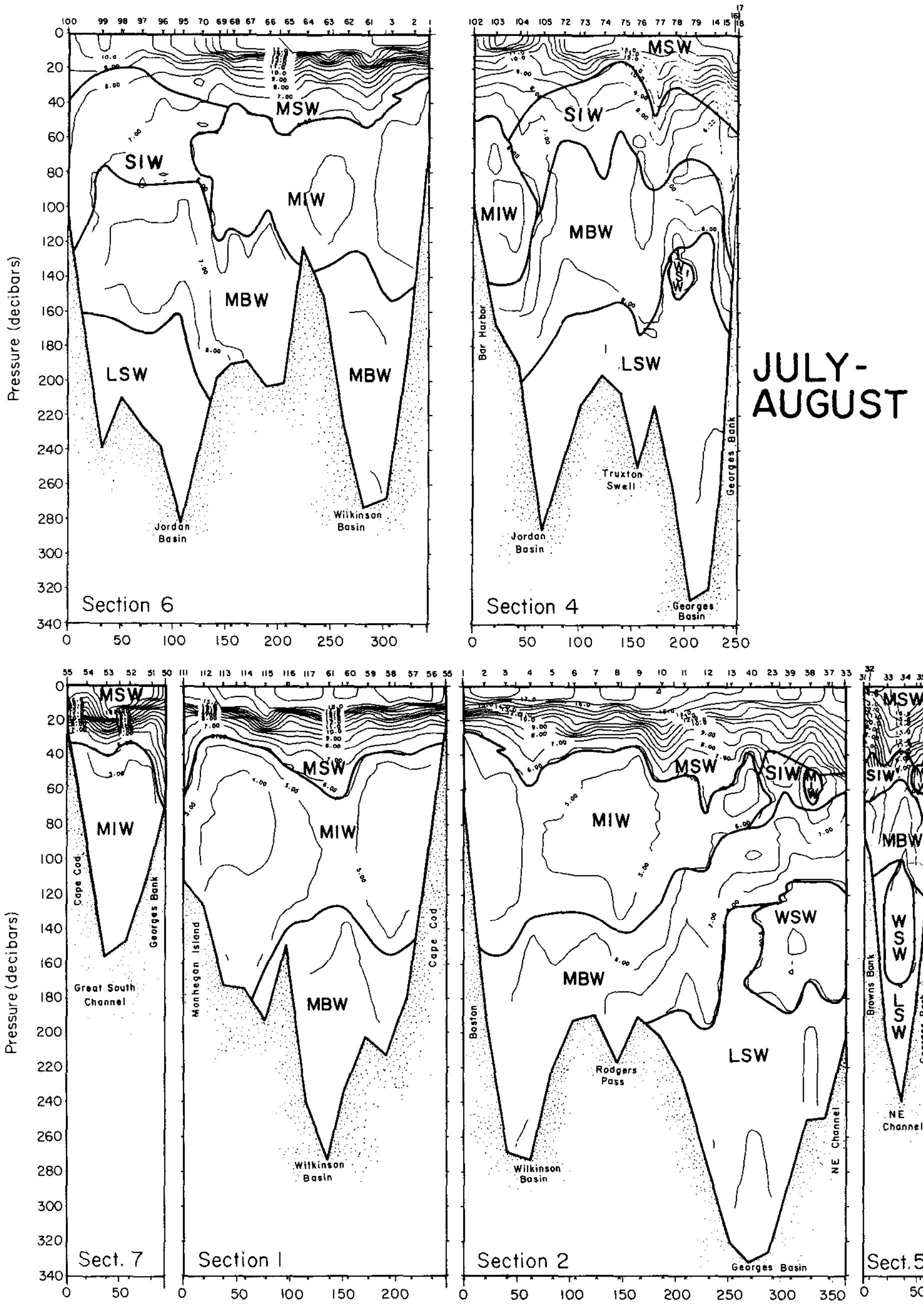


Figure 18. Water mass and temperature distribution sections for July–August 1987. See Figure 2 for section locations.

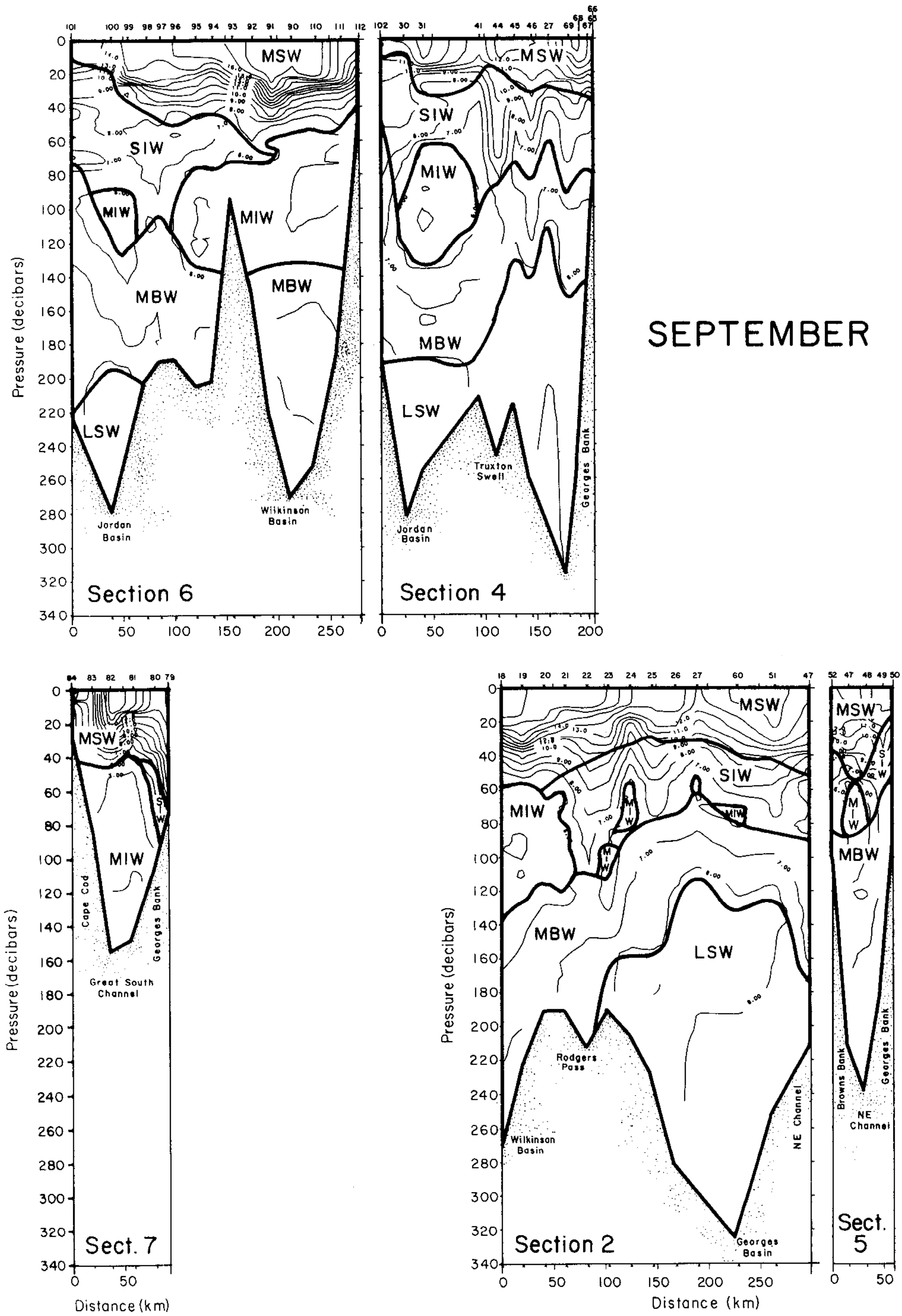


Figure 19. Water mass and temperature distribution sections for September 1987. See Figure 2 for section locations.

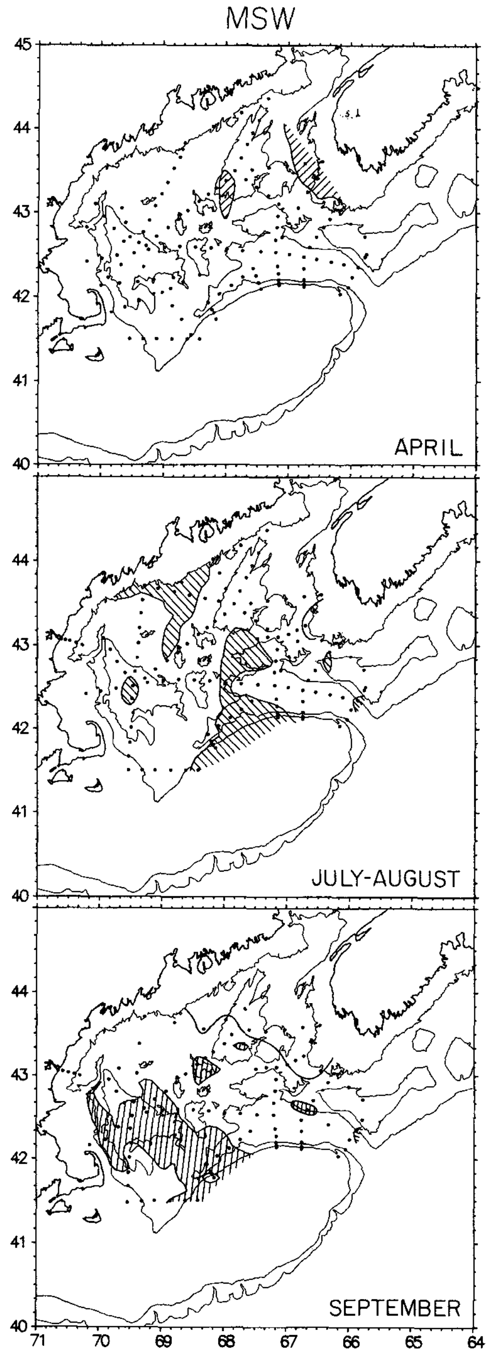


Figure 20. Evolution in the distribution of Maine Surface Water (MSW) thickness (m) from April through September 1987. The MSW thickness range between 0 m (bold contour) and 50 m is clear, while thicknesses in the range 50–100 m are hatched. In April, MSW is found only in the vicinity of the hatched regions, with Maine Intermediate Water actually reaching the surface throughout the rest of the Gulf (see text). The CTD station locations and the 100 m and 200 m isobaths are indicated.

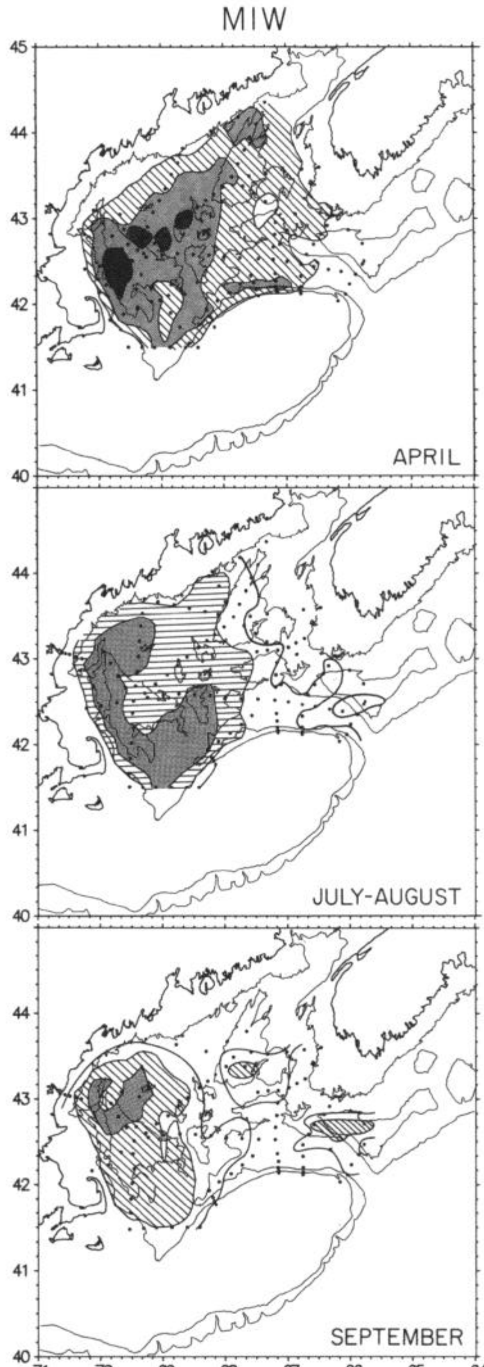


Figure 21. Evolution in the distribution of Maine Intermediate Water (MIW) thickness (m) from April through September 1987. The thicknesses in the April map reflect only MIW below 40 m depth (see text). The MIW thickness range between 0 m (bold contour) and 50 m is clear, while thicknesses in the range 50–100 m are hatched; 100–150 m stipled; and >150 m solid. The CTD station locations and the 100 m and 200 m isobaths are indicated.

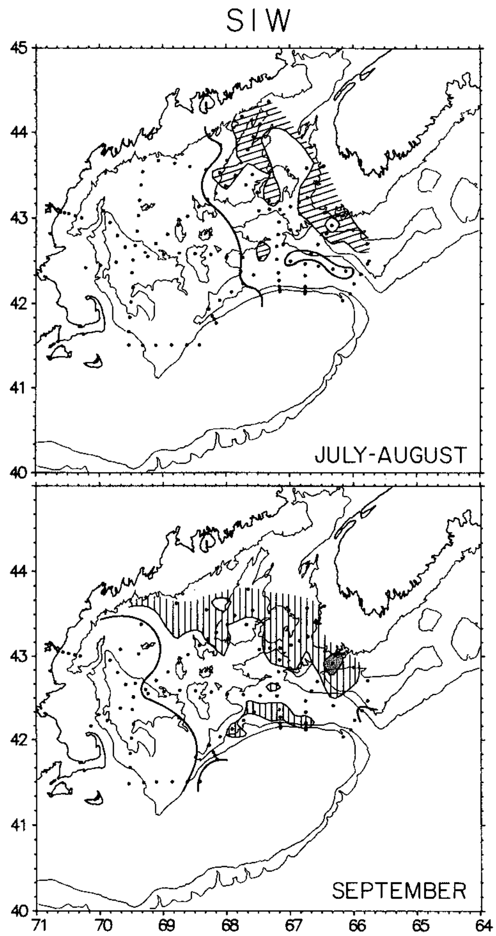


Figure 22. Evolution in the distribution of Summer Intermediate Water (SIW) thickness (m) from the July–August through September 1987 CTD surveys. The SIW thickness range between the 0 m (bold contour) and 50 m is clear, while thicknesses in the range of 50–100 m are hatched. The CTD station locations and the 100 m and 200 m isobaths are indicated.

The history of the normalized water mass volumes, as defined by the April, July/August and September 1987 survey observations (Fig. 26), features (a) a near-linear decrease of Maine Intermediate Water volume, (b) a near-linear increase of Summer Intermediate Water volume, and (c) a Slope Water volume maximum in August. Some of the decrease in Maine Intermediate Water resulted from outflow from the study region, with the rest lost through mixing. The increase of Slope Water (primarily Labrador) between April and August resulted from inflow through the Northeast Channel. An estimated inflow rate of 0.102×10^6 m^3 s^{-1}, based on the Slope Water volume increase alone, is a lower bound on the total inflow. Some additional Slope Water inflow was transformed (i.e., lost) through

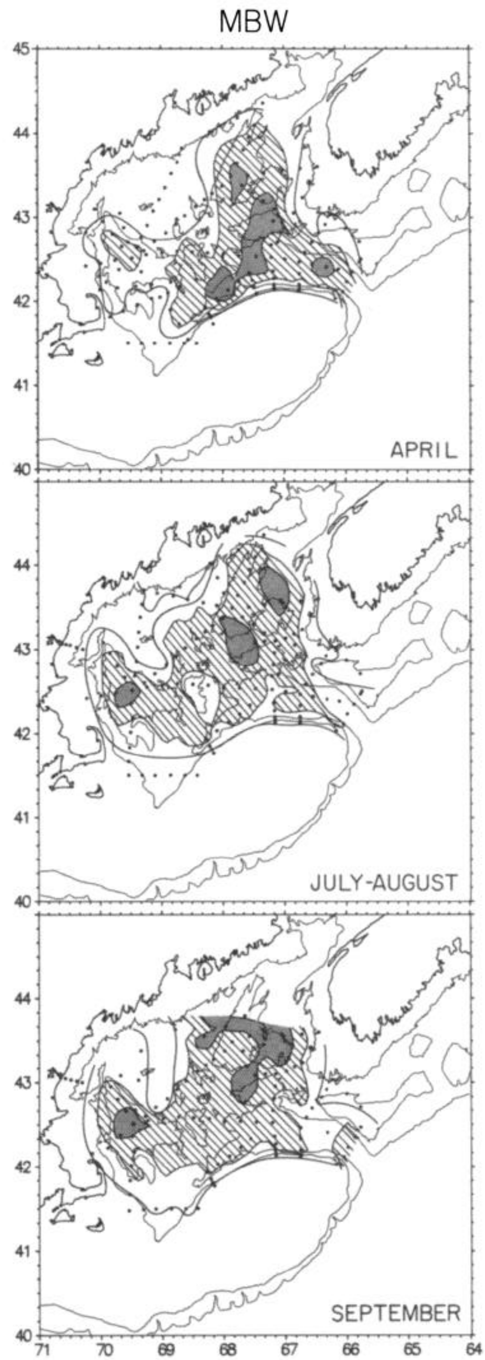


Figure 23. Evolution in the distribution of Maine Bottom Water (MBW) thickness (m) from April through September 1987. The MBW thickness range between 0 m (bold contour) and 50 m is clear, while thicknesses in the range 50–100 m are hatched, and 100–150 m stipled. The CTD station locations and the 100 m and 200 m isobaths are indicated.

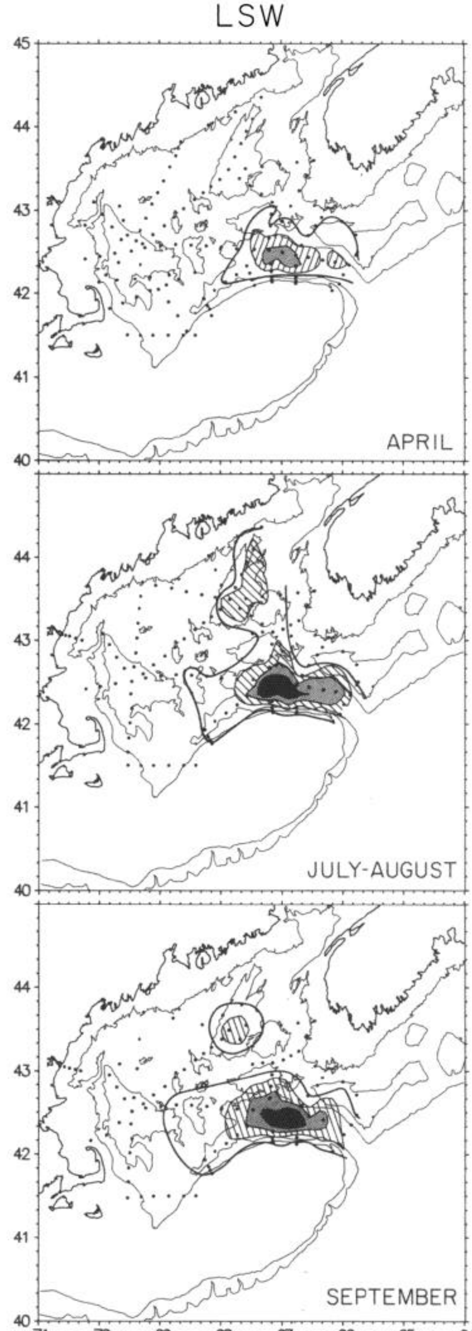


Figure 24. Evolution in the distribution of Labrador Slope Water (LSW) thickness (m) from April through September 1987. The LSW thickness range between 0 m (bold contour) and 50 m is clear, while thicknesses in the range 50–100 m are hatched; 100–150 m stipled; and >150 m solid. The CTD station locations and the 100 m and 200 m isobaths are indicated.

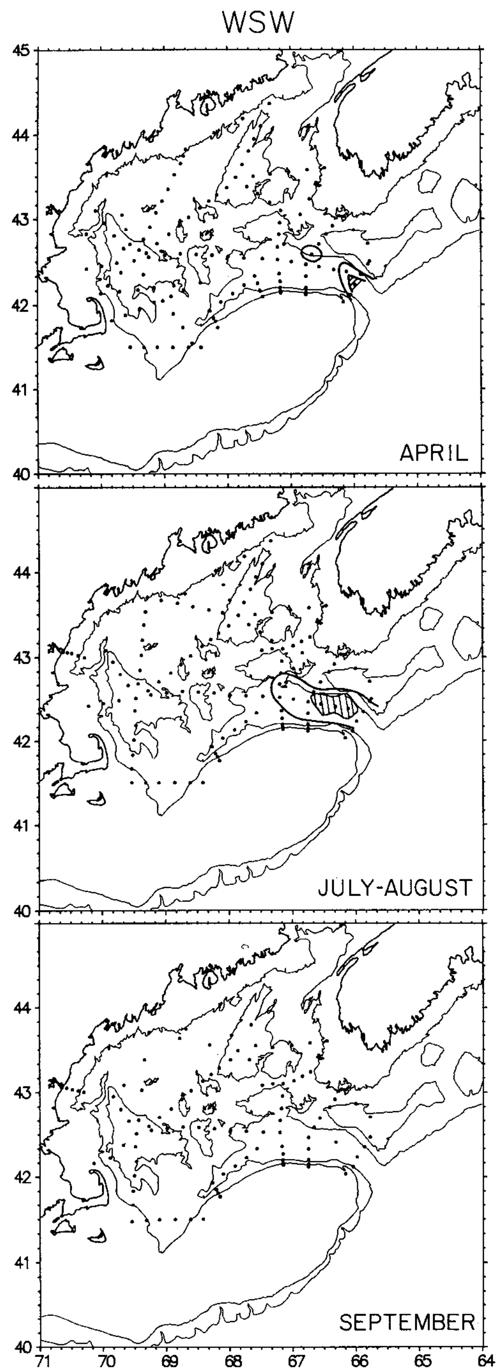


Figure 25. Evolution in the distribution of Warm Slope Water (WSW) thickness (m) from April through September 1987. The WSW thickness range between 0 m (bold contour) and 50 m is clear, while thicknesses in the 50–100 m range are hatched. The CTD station locations and the 100 m and 200 m isobaths are indicated.

Table 6. Normalized Gulf of Maine water mass volumes based on the April, July–August, and September 1987 CTD surveys. The volumes for Maine Surface Water (*MSW*), Maine Intermediate Water (*MIW*), Summer Intermediate Water (*SIW*), Maine Bottom Water (*MBW*), and Slope Water (*SW*) were normalized by the April volume so that the computed differences between surveys would be meaningful. The normalized and actual total volumes for each survey are given along with percent of total normalized volume for each water mass. The volume change rates are derived from the water mass volume differences.

	Normalized water mass volumes (10^{12} m^3)					Total volume (10^{12} m^3)		Total volume change rates (10^6 m^3/s)				
Survey	*MSW*	*SIW*	*MIW*	*MBW*	*SW*	Norm.	Actual	*MSW*	*SIW*	*MIW*	*MBW*	*SW*
April	2.95	—	6.42	3.13	0.59	13.09	13.09					
	22.5%	—	49.0%	23.9%	4.6%	100.0%						
								−0.035	+0.229	−0.289	−0.007	+0.102
Jul–Aug	2.62	2.18	3.67	3.06	1.56	13.09	15.84					
	20.0%	16.7%	28.0%	23.4%	11.9%	100.0%						
								−0.032	+0.226	−0.277	+0.121	−0.040
Sept	2.50	3.02	2.64	3.51	1.41	13.09	13.73					
	19.1%	23.1%	20.2%	26.8%	10.8%	100.0%						

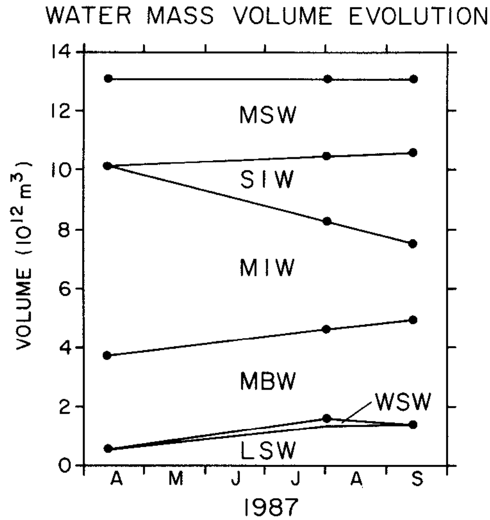


Figure 26. Gulf-wide water mass volume evolution between April and September 1987. The total volumes of each water mass, normalized according to the total April 1987 volume, were determined from the computations displayed in Figures 20–25. (Also see Table 6).

mixing with Intermediate Water to form Bottom Water. The fact that the volume of Bottom Water remained nearly constant between April and August suggests either a Bottom Water outflow and/or an insitu contribution to Summer Intermediate Water formation.

A simple water mass conservation model (Fig. 27) is used here to help explain the volume changes in the different water masses and in particular to assess the relative importance of inflow and local mixing in the evolution of water masses within the Gulf.

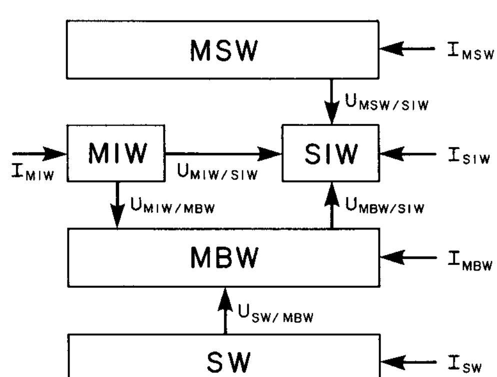


Figure 27. Schematic of water mass conservation model. The sense of the net inflow (I) and mixing (U) transports to/from each of the water mass domains is indicated. While net inflows can be negative, mixing transports cannot.

Conservation of total water mass volume within the observation domain is expressed as a balance of the *net inflows* of Maine Surface Water (*MSW*), Summer Intermediate Water (*SIW*), Maine Bottom Water (*MBW*), Maine Intermediate Water (*MIW*), and Slope Water (*SW*) masses, according to

$$I_{MSW} + I_{SIW} + I_{MBW} + I_{MIW} + I_{SW} = 0\,. \tag{1}$$

The presumption here is that water masses, with the defined Gulf water properties, can be formed around the periphery of the observation domain and then can enter the observation domain (i.e., the model Gulf). Of course, any of the water masses can also leave the model Gulf. In this model, the volumes of water masses within the Gulf change in time through either local mixing of adjacent water masses or net inflow (outflow) to (from) the Gulf. We assume that local surface heating/cooling and or evaporation/precipitation will change only surface water properties but not its volume.

The following outlines the detailed conservation statements of each water mass. For example, Slope Water volume V_{SW} is allowed to increase due to net inflow I_{SW} and decrease due to its contribution to the formation of Bottom Water (at a rate $U_{SW/MBW}$, via mixing with Maine Intermediate Water) within the Gulf according to

$$\frac{dV_{SW}}{dt} = I_{SW} - U_{SW/MBW}\,. \tag{2}$$

Since thermodynamics will not allow a particular water mass to *unmix* into its components, the signs of the intra-model mixing transports (*Us*) are positive definite. Bottom Water volume is allowed to increase due to both net inflow and Slope Water/Intermediate Water mixtures, and decrease due to contributions to Summer Intermediate Water formation within the model Gulf according to

$$\frac{dV_{MBW}}{dt} = I_{MBW} + U_{SW/MBW} + U_{MIW/MBW} - U_{MBW/SIW}\,. \tag{3}$$

Maine Intermediate Water volume is allowed to increase due to net inflow and decrease due to its contribution to both Bottom and Summer Intermediate Water formation within the Gulf according to

$$\frac{dV_{MIW}}{dt} = I_{MIW} - U_{MIW/SIW} - U_{MIW/MBW}\,. \tag{4}$$

It is important to note that we assume no significant mixing of Maine Intermediate and Maine Surface Water relative to other mixing possibilities within the observation domain—the Gulf. This restriction is rationalized by the relatively strong seasonal pycnocline between the water masses in the western Gulf. By contrast, Summer Intermediate Water volume is allowed to increase due to net inflow and through the mixing of Maine Intermediate, Bottom, and Surface Water in the more weakly stratified regions in the eastern Gulf and along the north flank of Georges Bank according to

$$\frac{dV_{SIW}}{dt} = I_{SIW} + U_{MIW/SIW} + U_{MBW/SIW} + U_{MSW/SIW}. \tag{5}$$

Surface Water volume is allowed to increase due to net inflow and decrease due to its contribution to Summer Intermediate Water production within the reduced seasonal pycnocline region in the eastern Gulf according to

$$\frac{dV_{MSW}}{dt} = I_{MSW} - U_{MSW/SIW}. \tag{6}$$

The mixing loss rates of Slope Water ($U_{SW/MBW}$) and Maine Intermediate Water ($U_{MIW/MBW}$), due to the production of Bottom Water, are assumed to be proportional to the rate change of Bottom Water volume, which has been corrected for (a) inflow *and* (b) its contribution to Summer Intermediate Water production according to

$$U_{SW/MBW} = \alpha_{SW/MBW}\left(\frac{dV_{MBW}}{dt} - I_{MBW} + U_{MBW/SIW}\right) \tag{7}$$

and

$$U_{MIW/MBW} = \alpha_{MIW/MBW}\left(\frac{dV_{MBW}}{dt} - I_{MBW} + U_{MBW/SIW}\right). \tag{8}$$

The α's in (7) and (8) indicate the proportion of the respective water types contributing to Bottom Water formation. For a specified mixing process, the α's sum to 1.00. Likewise, the mixing loss rates of Bottom Water, Maine Intermediate Water, and Surface Water to the production of Summer Intermediate Water, respectively, can be expressed according to

$$U_{MBW/SIW} = \alpha_{MBW/SIW}\left(\frac{dV_{SIW}}{dt} - I_{SIW}\right), \tag{9}$$

$$U_{MIW/SIW} = \alpha_{MIW/SIW}\left(\frac{dV_{SIW}}{dt} - I_{SIW}\right), \tag{10}$$

and

$$U_{MSW/SIW} = \alpha_{MSW/SIW}\left(\frac{dV_{SIW}}{dt} - I_{SIW}\right). \tag{11}$$

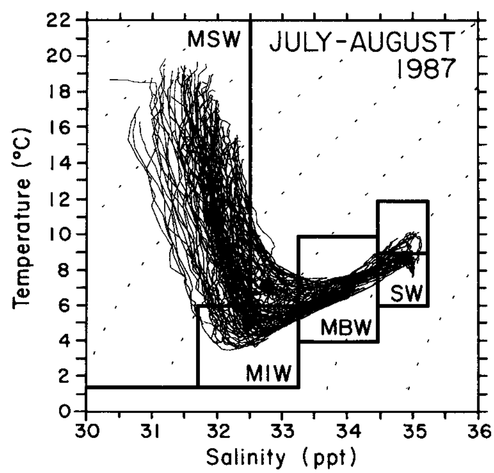


Figure 28. The assumed T-S characteristics for each of the model water masses are indicated by the solid dots superposed on the July–August T-S diagram.

The water mass volume rate changes within the Gulf are derived from observations (see Table 6), and the specific mixing proportions are estimated from T-S diagrams. Thus this system of Eqs. (1)–(11) reduces to four equations in the five unknown net inflow transports. The system is solved with the additional assumption that the Bottom Water inflow is proportional to Slope Water inflow according to

$$I_{MBW} = \gamma I_{SW}, \tag{12}$$

where γ is a specified constant of proportionality. The assumption is based on the observations that Bottom Water is always associated with Slope Water in the Northeast Channel (Figs. 17–19). We hypothesize that the mixing processes that form Bottom Water extend beyond our observation domain, enabling both inflow and outflow of Bottom Water. Based on the linkage between Bottom and Slope Water, we assume that their net inflows are proportional.

Approximations to the mixing coefficients were determined by assuming simple linear mixing between water types with representative temperature/salinity characteristics of the water masses of interest. We determined the characteristic temperature/salinity values for the different water masses involved from the July/August 1987 composite T-S diagram (Fig. 28)—namely, Surface Water (11°C/32 psu); Summer Intermediate Water (7°C/32.75 psu); Maine Intermediate Water (5°C/32.5 psu); Bottom Water (7°C/34 psu); and Slope Water (8°C/34.75 psu). Linear mixing between water types with these properties shows that (a) Slope Water and Maine Intermediate Water mix in proportions of approximately 2:1 to produce Bottom Water, and (b) Maine Intermediate Water, Bottom Water and Surface Water mix in proportions of approximately 2:1:1 to produce Summer Intermediate Water. Thus,

Table 7. Model transports (10^6 m^3 s^{-1}) for different γ's (see text).

	γ		
Transports	0.25	0.30	0.35
I_{SW}	0.107	0.104	0.109
I_{MBW}	0.027	0.031	0.038
I_{SIW}	0.066	0.063	0.010
I_{MIW}	−0.205	−0.205	−0.176
I_{MSW}	0.006	0.007	0.020
$U_{SW/MBW}$	0.005	0.002	0.007
$U_{MIW/MBW}$	0.002	0.001	0.003
$U_{MBW/SIW}$	0.041	0.042	0.055
$U_{MIW/SIW}$	0.082	0.083	0.109
$U_{MSW/SIW}$	0.041	0.042	0.055

the corresponding mixing coefficients are

$$\begin{aligned} \alpha_{SW/MBW} &= 0.667 \\ \alpha_{MIW/MBW} &= 0.333 \\ \alpha_{MIW/SIW} &= 0.500 \\ \alpha_{MBW/SIW} &= 0.250 \\ \alpha_{MSW/SIW} &= 0.250 \end{aligned} \tag{13}$$

The system of Eqs. (1)–(13) were solved for the April through August time period using the water mass volume rate change values from Table 6. The solution to this problem was very sensitive to the choice of γ—the coefficient linking Bottom Water and Slope Water net inflow. This sensitivity is related in part to the large contrast between the relatively small rate changes in the volumes of Surface Water and Bottom Water and large rate changes in the volumes of Maine Intermediate, Summer Intermediate and Slope Water.

Realistic solutions were obtained for the γ range $0.23 < \gamma < 0.37$. As indicated in Table 7, the solutions were not particularly sensitive to the choice of α in this range. Other values of γ were assumed in attempting to find model solutions. For $\gamma > 0.37$, $U_{SW/MBW}$ and other mixing transports were negative which is not allowed thermodynamically. At the other extreme of $\gamma = 0.10$, inflows of Slope, Bottom and Surface Water fed an unrealistically massive production of Summer Intermediate Water. This scenario was rejected as unrealistic because all Maine Intermediate Water decrease was involved in Summer Maine Intermediate Water decrease was involved in Summer Intermediate Water production, allowing none for the outflow observed by Flagg (1987) and others.

For the $\gamma = 0.25$ solution, appropriate for the April through July time period, 72% of the Maine Intermediate Water loss was due to export from the model Gulf domain. Maine Intermediate Water outflow was balanced primarily by a combined inflow of Slope Water, Summer Intermediate Water and Bottom Water. Of course, there is no way of determining how much of the "exported" Maine Intermediate water was subsequently mixed with Surface Water around the periphery of the Gulf—i.e., outside the observation domain. However, water mass distributions (Figs. 17–19) suggest that much of the Maine Intermediate Water was advected from the Gulf through the Northeast Channel. The Maine Intermediate Water mixing loss (28%) went into the local production of Summer Intermediate Water within the model Gulf.

The 72%–28% partitioning of export loss and mixing loss for Maine Intermediate Water is significantly different from the 40%–60% partition estimated by Hopkins and Garfield (1979) for summer 1966 observations (using a different approach). Short of applying the Hopkins and Garfield analysis scheme to the 1987 observations, there is no way to explain this large difference definitively. However, one possibility is that our model domain does not include all of the shallower regions where important mixing occurs. Thus some of our net export may actually be involved in mixing around the periphery of the Gulf. Another possibility is that the anomalously large amount of freshwater in the Gulf in 1987 may have inhibited mixing more than in other years. Resolving this very important difference is left to future work.

The model calculation suggests that 70% of the increase in Summer Intermediate Water volume was due to mixing within the model (i.e., observation) domain. The Summer Intermediate Water distributions (e.g., Fig. 22) show that this mixing occurred in the eastern Gulf and over the northeast flank of Georges Bank, where vertical stratification was relatively weaker. In this model scenario, 66% of the Bottom Water net inflow went into the Summer Intermediate Water production. Actually, there was very little *net* Bottom Water formed within the observation domain during the April–August period. In fact, only about 4% of the Slope Water inflow and 1% of the Maine Intermediate Water decrease went toward net Bottom Water production during the April–July time period.

Still, 30% of the increase in the Summer Intermediate Water volume within the model Gulf resulted from net inflow. As the Figure 22 distributions suggest, the Summer Intermediate Water inflow probably came from the eastern Gulf near Nova Scotia, and the northern flank of Georges Bank. Recent work by Tee *et al.* (1992) suggests that complicated tidal mixing and topographic upwelling processes in the region of Cape Sable and Browns Bank may be very important factors in producing Summer Intermediate Water. The Irish and Brown (1992) flow estimates for 1987 (Fig. 29) are qualitatively consistent with transporting the Summer Intermediate Water produced in the Cape Sable/Browns Bank region into our observation domain.

One of the weaknesses in the model is the ad hoc assumption as to the relation of Bottom and Slope Water inflow rate. A planned revisit of the Ramp *et al.* (1985) Northeast Channel observations will enable us to test the ad hoc assumptions related

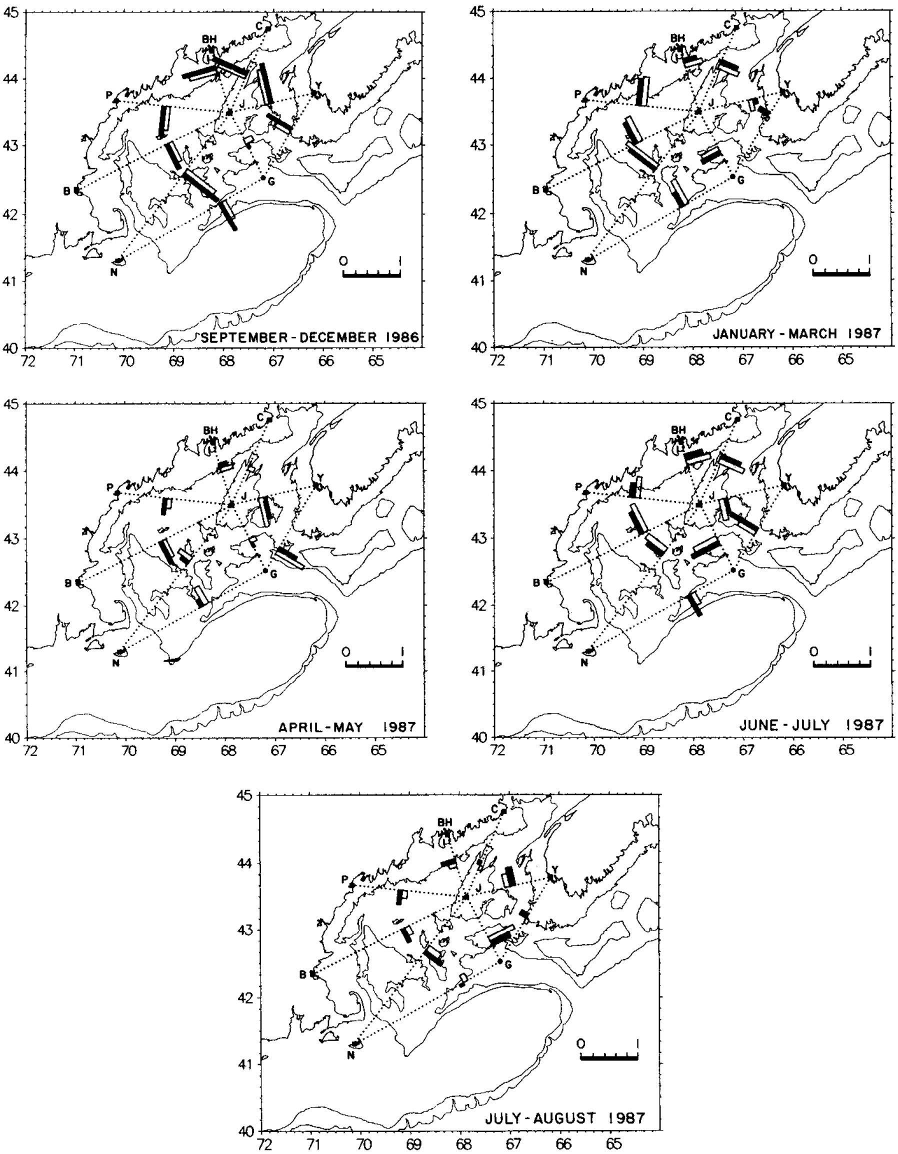


Figure 29. The mean geostrophic transport patterns in the Gulf of Maine. The flowthrough (11 September–31 December 1986), anticyclone (1 January–31 March 1987), outflow (11 July–14 August 1987) patterns are constructed from the temporal mean values (during the indicated periods) of the geostrophic transport time series (open bars) and the geostrophic transport adjusted by the Chapman *et al.* (1986) annual mean values (solid bars) (Brown and Irish, 1992) in units of 10^6 m^3/s.

to the use of and inferred value for γ. The in-depth exploration of many issues concerning Gulf of Maine circulation will require a comprehensive combined modeling/observation study with a particular focus on the Northeast Channel.

5. Gulf of Maine water mass histories: 1987

We are now in a position to summarize the principal features in spring/summer 1987 evolution of the principal water masses in the Gulf of Maine. The 1987 anomalies are highlighted.

a. Maine Surface Water. Surface Water in the 1987 Gulf of Maine (like any other year) exhibited the greatest range of properties of any of the Gulf of Maine water masses. Local river runoff and large seasonal changes in surface temperature are responsible for this wide range. In April 1987, most of the Surface Water above 40 m actually had the water property characteristics of Intermediate Water (Table 5). Still, there were a few patches of April Surface Water exceeding 50 m thickness in Jordan Basin and west of Yarmouth in April (Fig. 20). These patches—as cold but fresher than Intermediate Water—were clearly derived from the unusually large inflows of Scotian Shelf Water, as discussed above. The time-series T/S relations (Fig. 15) show that Surface Water in Jordan Basin warmed rapidly enough during May to produce the pycnocline that isolated Intermediate Water. We assume a similar timing for the separation of Surface Water and Intermediate Water in the western Gulf. The warming of Surface Water continued throughout the Gulf during June (Figs. 4 and 5). The thickness of the Surface Water layer in Jordan Basin increased from May through the middle of July, and then abruptly decreased.

The layer of Surface Water in Georges Basin was somewhat thinner than in the rest of the Gulf due to the presence of larger amounts of Slope Water (Fig. 16). The late May/early June inflow of Slope Water (see Fig. 29) was apparently accompanied by some Scotian Shelf Water with Surface Water properties (and other associated water masses). The inflow appears to be silhouetted in the July-August picture (Fig. 20) by a relatively thicker patch of Surface Water with apparent Georges Bank origins. The latter Surface Water patch probably consisted primarily of local spring river runoff that became entrained in the jet along the north flank of Georges Bank (see Brown and Irish, 1992). The other relatively fresh and thick patch of Surface Water, apparently being drawn offshore south of Bar Harbor, was probably fed by a combination of the spring river runoff to the northeastern Gulf and remnants of the earlier Scotian Shelf Water inflow.

In September, the thickest and freshest patches of Surface Water were found in the western Gulf. Surface geostrophic flow patterns (Brown and Irish, 1992) from this period explain the westward advection of these Surface Water patches from the eastern to the western Gulf.

b. Maine intermediate water. Most of the 1987 Intermediate Water was derived from winter water formed in and around Wilkinson and Jordan Basins during the winter of 1986–87. In April, the thickest layers (100 m–200 m) of Intermediate Water were in Wilkinson Basin (Figs. 17 and 21), while the thinnest layer was in Georges Basin (about 90 m). During May, the uppermost 40 m of Intermediate Water was converted to Maine Surface Water through warming. During June and July, most of Intermediate Water in Jordan and Georges Basin was replaced by Summer Intermediate Water (compare Figs. 17 and 18; 21 and 22) and the volume of Intermediate Water in Wilkinson Basin decreased.

What was the fate of the 1987 Maine Intermediate Water? The model results and direct observations suggest that about 30% was lost through mixing and 70% of the Maine Intermediate Water left the Gulf through the Northeast Channel or over Georges Bank. The presence of Maine Intermediate Water in the spring–summer 1987 Northeast Channel water mass sections (Fig. 17) and the sustained geostrophic outflow between Georges Basin and Yarmouth during April and May (Brown and Irish, 1992; Fig. 29) suggest the export route for Jordan Basin Intermediate Water. The presence of a relatively thick layer of Maine Intermediate Water in southern Wilkinson Basin in late July (Fig. 21) was consistent with the June–July average geostrophic transports in the western Gulf. However, deeply-drogued drifter observations from spring 1988 [Beardsley, personal communication] suggest that most of the water at 50 m depth in southern Wilkinson Basin flowed out along the north flank of Georges Bank. Thus we conclude that most of the 1987 Wilkinson Basin Intermediate Water left the Gulf via the Northeast Channel and/or over Georges Bank.

Between early August and mid-September, Intermediate Water disappeared completely from Georges Basin. The small amount left in Jordan Basin by September was concentrated in the southern sector—probably advected there from the northwest by the cyclonic circulation (Fig. 29). Most of the September Intermediate Water was located in Wilkinson Basin, with the largest concentrations in the northern part of the basin (Figs. 19 and 21). From that location, the Intermediate Water was in a position to supply cold water and nutrients to the coastal regions of the western Gulf during fall.

c. Summer Intermediate Water. According to our water mass definition scheme, there was no Summer Intermediate Water in the Gulf in April. After April, Summer Intermediate Water was introduced through inflow to and some mixing within the Gulf. In early summer, Summer Intermediate Water replaced Maine Intermediate Water at mid-depth in the eastern Gulf of Maine—earlier in Georges Basin and later in Jordan Basin (Figs. 16, 18, and 19). Summer Intermediate Water had the properties of an *in situ* mixture of Bottom, Maine Intermediate and a warming Surface Water (see Fig. 28); suggesting mixing within the eastern Gulf observation domain. The distribution maps (Fig. 22) suggest, that Summer Intermediate Water was also advected into the eastern Gulf from regions near Nova Scotia. This is consistent with the fact that the June/July geostrophic inflow through the Georges/ Yarmouth section (Fig. 29) included Summer Intermediate Water. Thus it appears that during spring and early summer 1987 Summer Intermediate Water was formed through a complicated mixing of Surface, Maine Intermediate, and Bottom Water in the eastern Gulf of Maine. The general cyclonic circulation that developed in Jordan Basin in August then helped to draw the Summer Intermediate Water from the northeastern Gulf into the north-central Gulf by September.

d. Maine Bottom Water and Slope Water. Bottom Water distribution is linked rather closely to Slope Water distribution. The Bottom Water layer in the Gulf—typically 80 m thick—is derived primarily from the mixing of both Labrador and Warm Slope Water with the overlying Maine and Summer Intermediate Water masses. During spring 1987, the mean depth of the Bottom Water layer, varied considerably—from 220 m in Jordan Basin to 45 m in central Georges Basin during early summer (see Fig. 16). The variability in the mean depth of Bottom Water seems to have been linked to the amount of Slope Water in the Gulf.

In April 1987 there were significant amounts of Labrador Slope Water in the Northeast Channel and Georges Basin, but none in Jordan or Wilkinson Basin (Fig. 24). There was geostrophic inflow to the Gulf through the Georges Basin/ Yarmouth section between September 1986 and March 1987 (Brown and Irish, 1992; Fig. 29). Apparently the inflow of Slope Water during this period was just sufficient (a) to keep Georges Basin filled with Labrador Slope Water to the depth of the sill and (b) to maintain the Bottom Water requirement elsewhere. There were signs of increased inflows of Slope Water during April. A series of pulses of Warm Slope Water surrounded by Labrador Slope Water (see Fig. 16) increased the amount of Slope Water in the Northeast Channel and Georges Basin between April and June (Figs. 17 and 18; 24 and 25).

In April, the greatest concentration of Bottom Water was in the eastern Gulf, particularly Georges Basin. [Note that the water mass thickness time series (Fig. 16) were misleading in showing nearly equal thicknesses of Bottom Water in the centers of both Georges and Jordan Basins.] This early spring scenario was consistent with the notion that significant amounts of Bottom Water were entrained in the production of winter water in the western Gulf. Note how the distributions of Intermediate and Bottom Water in April complemented each other (Figs. 21 and 23).

Between April and August, the combined inflow of Slope Water and interior circulation patterns distributed the Bottom Water more uniformly throughout the Gulf. The early August distributions of Bottom and Labrador Slope Water (Figs. 23 and 24) were consistent with the variable geostrophic flow scenarios reported by Brown and Irish (1992) for the April–August time period.

6. Final comments

This paper focuses on the formation and evolution of Maine Intermediate Water between April/May, when Maine Surface and Intermediate Water masses become distinct, and the end of the summer. The Maine Intermediate Water volume history is probably one of the more sensitive indications of the seasonal and interannual changes in the Gulf of Maine water properties. While direct comparisons between 1987 and other years are difficult because of the quasi-subjectivity in the water mass definitions, it appears that unusually large amounts of Maine Intermediate Water were produced in 1987 (Table 4). Maine Intermediate Water filled about half of the observed Gulf in May and one fifth even in September. This large percentage of Maine Intermediate Water was consistent with the relatively deep convective mixing in winter 1987 that was apparently responsible for making the 1987 Gulf colder than the 1986 Gulf. The anomalously large inflows of freshwater from the Scotian shelf also were responsible for making the 1987 Gulf fresher than the 1986 Gulf. It is not clear how the seemingly separate set of meterologically induced situations are linked climatologically. Those studies are likely to be done soon.

There was a net summertime loss of about 60% of the original 1987 Maine Intermediate Water—about 70% by export and 30% by mixing. That loss was balanced primarily by the inflow of Slope Water and the replacement by and/or conversion to somewhat warmer Summer Intermediate Water. The Summer Intermediate Water appears to have been formed in the far eastern Gulf. Because the replacement of Maine Intermediate by Summer Intermediate Water was nearly isopycnal, there were not significant circulation changes associated with these changes. Instead, it was the inflows and advection of buoyant patches of Maine Surface Water and dense patches of Slope/Bottom Water that were primarily responsible for the spring/summer evolution of the mesoscale Gulf circulation patterns as described by Brown and Irish (1992). The latter issues will be addressed in an upcoming paper.

7. Summary

The 1986–87 annual variation in the structure and disposition of the principal water masses in the Gulf of Maine has been described using an integrated set of quasi-synoptic and time series water property observations. The time series observations document the autumn preconditioning of upper water column, and the convection-induced mixed-layer deepening during the winter. The 1987 inflow of unusually large amounts freshwater from the Scotian Shelf inhibited vertical mixing in the eastern Gulf much more than in the western Gulf. Deep convective mixing formed the largest amounts of winter water in and around Wilkinson Basin. The 1987 Gulf was cooler than the 1986 Gulf because of the size of the 1987 Maine Intermediate Water mass.

The freshening and warming of the surface layers during April and May marked the beginning for a set of senarios spring/summer evolution and redistribution of all of the water masses in the Gulf. Definitions of the principal water masses in the Gulf—namely, Maine Surface Water, Maine Intermediate Water, and Maine Bottom Water—were based on April 1987 CTD survey information and historical definitions of Slope Water. The May-through-July time series observations of water mass thickness at the mooring sites revealed (1) the progression of Slope and Bottom Water from the Northeast Channel to Georges Basin and on to Jordan Basin and (2) the replacement of Maine Intermediate Water with Summer Intermediate Water—a newly defined water mass with eastern Gulf origins.

An April through September 1987 sequence of CTD-derived water mass structure maps document (1) the decrease in Intermediate Water volume from 49% of the observed Gulf in April to 20% in September; (2) the replacement of Intermediate Water by westward spreading of Summer Intermediate Water in the eastern Gulf; and (3) evolving structure of the inflowing Slope Water and associated Bottom Water. A simple water mass conservation model suggests that (a) a combination of Slope Water inflow (0.11×10^6 m^3/s), Bottom Water inflow (0.03×10^6 m^3/s), and Summer Intermediate Water inflow (0.07×10^6 m^3/s) balanced the Maine Intermediate Water outflow; (b) 71% of the Maine Intermediate Water loss was exported from the observation domain at a rate of 0.21×10^6 m^3/s; and (c) the rest of the Maine Intermediate Water loss was involved in the local production of Summer Intermediate Water through mixing.

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