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Interannual variability of boundary fluxes and water mass properties in the Gulf of Maine and on Georges Bank: 1993–1997^{☆,☆☆}

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Abstract

Analysis of three years (October 1993–September 1996) of monthly mean current, temperature and salinity observations from moorings in the major Gulf of Maine (GOM) inflows off southwest Nova Scotia (C2) and in Northeast Channel (NECE) reveals some new features of the annual NECE cycles, including (1) a peak in near-surface (< 75 m) inflow in spring (vs. late summer at depth), suggesting that dynamic control of shallow and deep layers may be different, (2) a maximum near-surface cross-channel flow toward Georges Bank (GB) in late winter, suggesting a climatological tendency for Scotian Shelf “cross-overs” in that season, and (3) the absence of a significant salinity cycle over most of the water column. Deviations from the annual cycle indicate that the first part of the observation period was characterized by enhanced warm, saline, deep (> 75 m) NECE inflow, followed by later episodes of enhanced cold, fresh inflow at C2 and shallow NECE. Generally, the flow rates at C2 and deep NECE were out of phase, with increased inflow at C2 (deep NECE) associated with reduced inflow at deep NECE (C2) and cooler, fresher (warmer, saltier) conditions at both sites. Freshwater transport anomalies to the Gulf are maximum in the surface layers and largely negative(positive) over the first(last) half of the measurements. The timing of these freshwater inflow variations is consistent with observed fluctuations in hydrographic measurements in the GOM and GB, which reached peak salinities in late 1994, then declined through 1995–1996. Oxygen isotope analysis suggests that almost all of the fresh water present on the central cap of GB in 1996 and early 1997 is of northern (Scotian Shelf)

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origin as opposed to 1994 and 1995 when Maine River Waters contributed 38 and 26%, respectively, to the freshwater (relative to 34.8) on the cap. A simple box model driven by observed changes in the boundary fluxes indicates that over the last half of the measurement period (April 1995–September 1996), the volumetric flow rate through the GOM increased by $10^5 \text{ m}^3 \text{ s}^{-1}$ (roughly 17% of the total transport, $5.83 \times 10^5 \text{ m}^3 \text{ s}^{-1}$), and that increased freshwater fluxes in the surface layers at C2 and NECE produced a net decrease of 0.73 in the salinity of the outflow waters. Average volumetric transports at C2(NECE) were roughly twice(half) those observed in the late 1970s, but the total is consistent with climatological estimates. The net *change* in the freshwater fluxes exceeds the total climatological mean estimate. Examination of possible local and remote sources confirms that the origin of the 1996–1997 freshwater anomaly is in the northern Labrador Sea/Baffin Bay and results from exceptionally cold winters in the early 1990s. Analysis of a similar event in the early 1980s suggests their occurrence is part of a quasi-decadal climate signal which follows the North Atlantic Oscillation (NAO). Crown copyright © 2000 Published by Elsevier Science Ltd. All rights reserved.

1. Introduction

One of the central goals of the US GLOBEC Georges Bank program is to gather and interpret information related to forecasting the effects of climate change on the Bank's productive ecosystem. As a proxy for climate change, the observed biophysical interannual variability is to be used to estimate the sensitivity to environmental changes and then to project the impact of expected shifts in the climate regime. A full understanding of the process also requires discovery of the sources of the observed variability and description of the dominant mechanism(s) involved. Recent work by Petrie and Drinkwater (1993) has examined decadal variability (e.g., cold 1960s vs. warm 1970–1980s) on the Scotian Shelf/Gulf of Maine and provided evidence that the primary cause of the dramatic cooling in the 1960s was an increase in the westward transport of the Labrador Current into the slope water region. The decadal signal was coherent on a large (regional) scale, with maximum range (4.6°C for temperature, 0.7 for salinity) occurring at the depth of 100 m over the continental slope. The anomalous waters penetrated the inner basins of the shelf through deep channels (e.g., Northeast Channel) and mixed upward into the surface layers. In contrast, we are concerned here with interannual variability on shorter time scales of 1–3 yr. Our goals are: (1) to describe, with reference to earlier observations, the variability of boundary fluxes and water mass properties in the Gulf of Maine/Georges Bank system during the GLOBEC field years (1994–1996), and (2) to investigate the sources of this variability.

The primary advective sources of water to the Gulf are thought to be the inflow of fresher surface water (SSW) from the Scotian Shelf and the deep inflow of slope water through Northeast Channel (NEC). These inputs, together with local river runoff and surface fluxes of heat and salt (E–P), control the water properties of the interior Gulf and Georges Bank (Bigelow, 1927). Smith (1983) estimated the annual mean volumetric transport of SSW past Cape Sable, N.S., at $140 \times 10^3 \text{ m}^3 \text{ s}^{-1}$ and identified strong seasonal cycles in the current, temperature and salinity such that the maximum inflow in winter was nearly double the mean, while the summer minimum was near zero. Furthermore, the presence of a clockwise gyre on the western cap of Browns Bank suggested that most of the net along-isobath transport was concentrated near the coast, i.e. within the 100 m isobath. Later, Smith (1989b) used multiyear time series from two long-term moorings off

Cape Sable to investigate interannual variability of the SSW inflow properties. He found that atmospheric forcing and warm core rings offshore may influence heat and salt flux anomalies, but that St. Lawrence River runoff anomalies had no significant effect.

Mountain and Manning (1994) explored the annual and interannual variability of temperature, salinity and density structure of the surface layer (0–50 m) in the Gulf using hydrographic data from the Marine Resources Monitoring, Assessment and Prediction (MARMAP) surveys (1977–1987). They found that the annual cycle of temperature, with a March minimum and range of 8–12°C, followed the surface heating/cooling cycle, but the salinity cycle (range: 0.6–1.0) was dominated by SSW inflow in the east and local runoff in the west. Thus the phase of the annual salinity minimum progressed counterclockwise around the Gulf, reaching Georges Bank in late summer. Interannual variability of the surface layer temperature was weak (1–2°C) in comparison to the annual cycle; but fluctuations of salinity (0.5), resulting from changes in local river input, surface fluxes, and, to a lesser extent, St. Lawrence River discharge, were nearly as large as the amplitude of the annual cycle. One puzzling result of this analysis is the lack of significant correlations between interannual surface salinity anomalies in the inflow region of Cape Sable and those in the interior of the Gulf, which the authors attributed to dominance of local river input and precipitation processes. The weak interannual temperature anomalies were generally not well-correlated over the region.

Ramp et al. (1985) estimated the annual mean inflow of deep (> 75 m) slope water through Northeast Channel at $260 \pm 60 \times 10^3 \text{ m}^3 \text{ s}^{-1}$ and an annual cycle such that the peak inflow of $350 \times 10^3 \text{ m}^3 \text{ s}^{-1}$ occurs in late summer. In winter, the inflow consists of intermittent “bursts” of axial flow that are strongly correlated with long- and cross-shore wind components and the subsurface pressure field in the Gulf. The stronger, steadier summer inflow is consistent with deep water properties detected by Mountain and Jessen (1987) in the interior basins of the Gulf with MARMAP hydrographic data (1978–1983). In the deep basins, weak annual cycles are characterized by May–June minima caused by downward mixing of cold intermediate water, followed by gradually increasing T, S through December due to advection of the NEC deep inflow. Also, mean T, S decrease in a counterclockwise sense through the inner basins (Georges, Jordan, Wilkinson) of the Gulf, i.e. following the circulation. Long time series of deep basin properties show interannual variability with ranges of 2°C and 0.4, attributed to similar fluctuations in the SSW and NEC inflows. Besides the direct effect of freshwater inflow on the Gulf water properties, large inflows have the secondary effect of reducing surface salinities in the western Gulf to the extent that deep winter mixing is inhibited (Mountain and Jessen, 1987). This, in turn, would influence the properties of near-surface waters flowing onto Georges Bank.

Questions concerning the origin of Georges Bank water cannot be answered unambiguously with T, S alone, because of the multiplicity of sources and processes affecting heat and freshwater content. This problem can be alleviated, to some extent, by knowing the relative depletion of the oxygen isotope, ^{18}O , a conservative tracer in seawater, because of the strong latitudinal dependence of that quantity in meteoric waters and hence freshwater runoff (Fairbanks, 1982). Chapman et al. (1986) and Chapman and Beardsley (1989) have used limited amounts of oxygen isotope data to hypothesize that the primary source of freshwater to northeast US coastal waters is located in the northern Labrador Sea. In the present context, we use the same oxygen isotope tracer to discriminate between local and remote sources of Georges Bank freshwater and their interannual variability.

2. Data and methods

2.1. Boundary fluxes

Long-term moored measurements of current, temperature and salinity were made at three primary sites (C2, NECE, NECW; Fig. 1a) during the 3-yr GLOBEC Phase I field effort (October 1993–September 1996). (For the period June 1994–June 1995, the NECW mooring was replaced by the NEP mooring in support of the 1995 Stratification Study.) Attention here is focussed on the primary inflow moorings: C2 on the 110 m isobath off Cape Sable, N.S., and NECE on the 210 m isobath on the eastern side of Northeast Channel.

The C2 mooring carried three current meters located near the surface (20 m), near the bottom (10 m above), and at mid-depth (50 m); NECE carried near-surface (20 m), near-bottom (20 m above) plus three current meters distributed over the water column (50, 100, 150 m). Near-surface instruments were isolated on separate moorings from those at 50 m and below to protect the latter from surface wave pumping. The subsurface floats on these two moorings were located just above the shallowest instruments, i.e. at 19 and 49 m. Instruments were usually Aanderaa RCM-7 or -8 current meters, sampling at 1-h intervals with estimated accuracies of speed ($\pm 1.3\%$), direction ($\pm 5^\circ$), temperature ($\pm 0.04^\circ\text{C}$), and salinity (± 0.1), based on laboratory and field calibrations. All instruments from the program were recovered, owing in large part to the placement of three guard buoys in a triangular array around each mooring pair. Overall data return was in excess of 90%, with all losses occurring in the salinity variable due to biological fouling of the conductivity sensor.

The current measurements were resolved into along-isobath ($+V$ at 284°T for C2, 305°T for NECE) and cross-isobath ($+U$ at 14°T , 35°T for C2, NECE) components and all records were filtered (half-power point at 31 h) to remove tides and other high-frequency signals. To quantify the seasonal variations, all variables were then averaged by calendar month and fit with a mean plus annual cycle using a least-squares multiple regression technique with $n - 3$ degrees of freedom, where n (≤ 36) is the number of months with good data (see Smith, 1989b). The estimator was of the form

$$Y_v = B_0 + B_1 \cos[\pi(t_v - \Phi_1)/6] + \varepsilon,$$

where t_v is the month number ($= 0.5$ for January, $v = 1$) and ε is the residual. The coefficients for mean (B_0) and statistically significant annual cycles ($B_1 = \text{amplitude}$, $\Phi_1 = \text{phase in months}$) are presented in Table 1(a), with the associated 95% confidence limits, (Δ , δ), based on Butman and Beardsley (1987).

The local depth-integrated monthly mean transport functions for volume (q), fresh water (b), and temperature (f) are estimated using the monthly mean values of V , S , and T in:

$$q = \int_{-H}^0 V \, dz \approx \sum_{i=1}^N V_i \Delta Z_i, \quad (1)$$

$$b = \frac{1}{S_0} \int_{-H}^0 V(S_0 - S) \, dz = q - \frac{1}{S_0} \int_{-H}^0 VS \, dz \approx q - \frac{1}{S_0} \sum_{i=1}^N V_i S_i \Delta Z_i \quad (2)$$

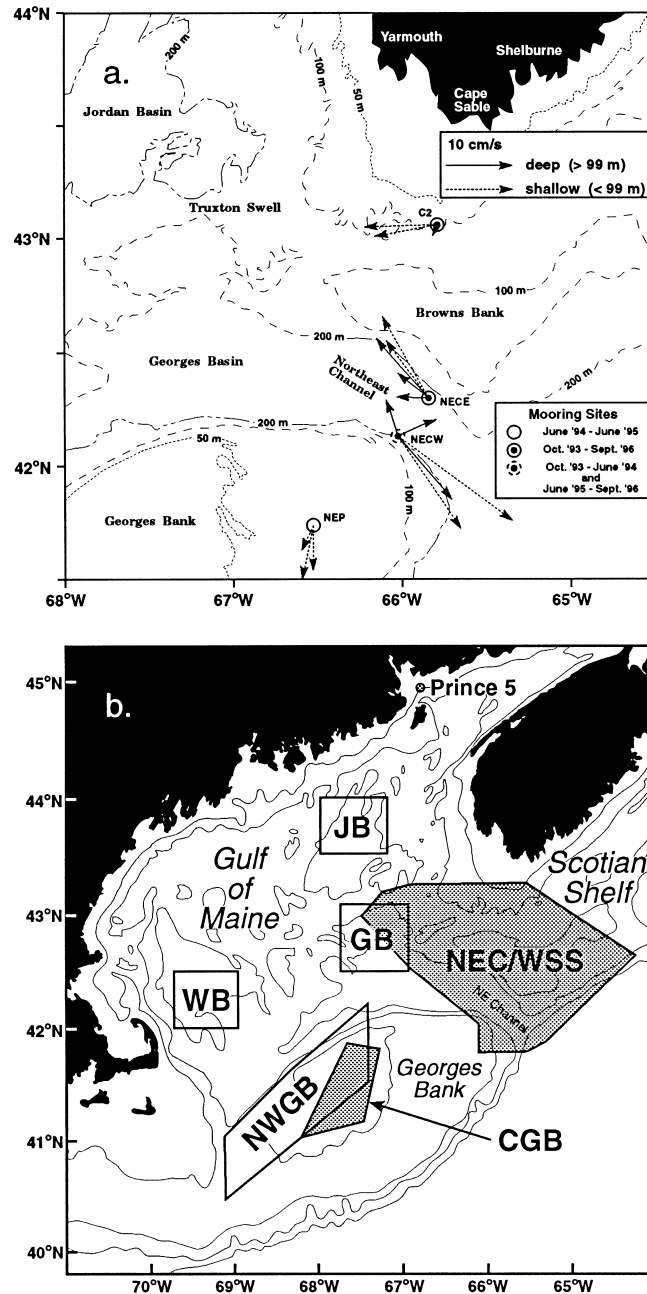


Fig. 1. (a) GLOBEC long-term mooring sites in Northeast Channel (NECE NECW) and off Cape Sable N.S. (C2). The NECW mooring was replaced by NEP during June 1994–June 1995. Vectors indicate record mean currents derived from the monthly means (see Table 1(a)), (b) Sampling zones for NEFSC standard hydrographic stations (open) and oxygen isotope ratios (hatched) in the Gulf of Maine.

Table 1

Variable (units)	Mean		Annual		$\sigma_{\text{R}}^{\text{a}}$
	n	B_0	$B_1 \pm \Delta_1^{\text{a}}$	$\Phi_1 \pm \delta_1^{\text{a}}$	
<i>(a) Significant seasonal cycles in boundary flux measured variables (U, V, T, S)</i>					
C2(26) – $U(\text{cm s}^{-1})$	36	– 4.0	2.2 ± 1.8	8.4 ± 1.6	3.8
$V(\text{cm s}^{-1})$	36	8.2	3.5 ± 2.3	6.6 ± 1.3	4.9
$T(^{\circ}\text{C})$	36	5.8	3.9 ± 0.6	8.8 ± 0.3	1.2
S	33	31.7	0.4 ± 0.2	7.8 ± 0.9	0.4
C2(50) – $U(\text{cm s}^{-1})$	36	– 3.1	1.7 ± 1.4	8.6 ± 1.5	2.9
$V(\text{cm s}^{-1})$	36	9.7	3.6 ± 2.1	6.3 ± 1.1	4.3
$T(^{\circ}\text{C})$	36	5.3	2.8 ± 0.6	9.0 ± 0.4	1.2
S	36	32.1	0.4 ± 0.2	8.0 ± 0.8	0.4
C2(100) – $U(\text{cm s}^{-1})$	36	– 1.4	— $\pm$ — ^b	— $\pm$ —	3.3
$V(\text{cm s}^{-1})$	36	0.6	— $\pm$ —	— $\pm$ —	1.9
$T(^{\circ}\text{C})$	36	5.2	1.7 ± 0.7	9.6 ± 0.7	1.4
S	36	32.6	0.4 ± 0.2	9.1 ± 0.7	0.3
NECE(22) – $U(\text{cm s}^{-1})$	36	3.0	5.2 ± 1.5	7.2 ± 0.5	3.1
$V(\text{cm s}^{-1})$	36	10.2	7.2 ± 3.0	3.3 ± 0.8	6.2
$T(^{\circ}\text{C})$	36	8.0	4.5 ± 0.8	8.6 ± 0.3	1.7
S	36	32.5	— $\pm$ —	— $\pm$ —	0.7
NECE(50) – $U(\text{cm s}^{-1})$	36	5.1	3.8 ± 1.1	8.0 ± 0.6	2.3
$V(\text{cm s}^{-1})$	36	12.8	3.7 ± 2.3	5.0 ± 1.2	4.9
$T(^{\circ}\text{C})$	36	7.8	1.9 ± 0.9	9.5 ± 0.9	1.9
S	36	33.3	— $\pm$ —	— $\pm$ —	0.5
NECE(100) – $U(\text{cm s}^{-1})$	36	2.4	3.0 ± 1.2	8.2 ± 0.6	2.5
$V(\text{cm s}^{-1})$	36	10.8	3.0 ± 1.9	7.1 ± 1.2	4.0
$T(^{\circ}\text{C})$	36	9.5	0.8 ± 0.7	11.1 ± 1.7	1.5
S	36	34.4	— $\pm$ —	— $\pm$ —	0.4
NECE(150) – $U(\text{cm s}^{-1})$	36	– 0.1	0.8 ± 0.7	2.3 ± 1.7	1.4
$V(\text{cm s}^{-1})$	36	5.8	2.1 ± 1.8	6.4 ± 1.7	3.8
$T(^{\circ}\text{C})$	36	10.2	0.7 ± 0.4	11.8 ± 1.1	0.9
S	36	35.3	— $\pm$ —	— $\pm$ —	0.2
NECE(190) – $U(\text{cm s}^{-1})$	36	– 2.7	0.9 ± 0.6	7.7 ± 1.4	1.3
$V(\text{cm s}^{-1})$	36	4.1	2.5 ± 1.4	5.5 ± 1.1	3.0
$T(^{\circ}\text{C})$	36	9.1	0.6 ± 0.2	0.6 ± 0.7	0.5
S	36	35.2	0.1 ± 0.1	11.7 ± 1.5	0.1
<i>(b) Significant seasonal cycles in boundary flux estimates (q, b, f)</i>					
C2 – $q_2 (\text{m}^2 \text{s}^{-1})$	36	7.5	3.1 ± 1.8	0.4 ± 1.2	3.8
$10^3 b_2 (\text{m}^2 \text{s}^{-1})$	36	632.0	322.0 ± 18.0	0.7 ± 1.1	374.0
$f_2 (\text{m}^2 \text{s}^{-1} ^{\circ}\text{C})$	36	– 22.2	26.3 ± 7.7	8.3 ± 0.6	16.1
NECE – $q_{\text{T}} (\text{m}^2 \text{s}^{-1})$	36	18.5	5.4 ± 3.2	5.2 ± 1.1	6.7
$10^3 b_{\text{T}} (\text{m}^2 \text{s}^{-1})$	36	447.0	191.0 ± 129.0	3.8 ± 1.3	269.0
$f_{\text{T}} (\text{m}^2 \text{s}^{-1} ^{\circ}\text{C})$	36	11.4	27.4 ± 13.9	9.1 ± 1.0	29.1

Table 1 (continued)

Variable (units)	Mean		Annual		σ_R^a
	n	B_0	$B_1 \pm \Delta_1^a$	$\Phi_1 \pm \delta_1^a$	
Shal – q_S ($m^2 s^{-1}$) ^c	36	8.8	3.5 ± 1.9	4.1 ± 1.1	3.9
$10^3 b_S$ ($m^2 s^{-1}$)	36	455.0	196.0 ± 113.0	3.8 ± 1.1	236.0
f_S ($m^2 s^{-1} ^\circ C$)	36	– 6.3	23.8 ± 7.9	9.0 ± 0.6	16.4
Deep – q_D ($m^2 s^{-1}$) ^c	36	9.7	3.2 ± 2.2	6.5 ± 1.3	4.6
$10^3 b_D$ ($m^2 s^{-1}$)	36	– 8.0	— $\pm$ — ^b	— $\pm$ —	84.0
f_D ($m^2 s^{-1} ^\circ C$)	36	17.6	— $\pm$ —	— $\pm$ —	15.3

^a Annual cycles are expressed as amplitude and phase in months measured from the start of the year; (Δ, δ) are the estimated 95% confidence intervals (see text); σ_R = standard deviation of the residuals.

^b Missing values indicate insignificant amplitudes at 95% confidence level.

^c Shal = shallow fluxes (≤ 75 m) at NECE; Deep = deep fluxes (≥ 75 m) at NECE.

and

$$f = \int_{-H}^0 V(T - T_0) dz \approx \sum_{i=1}^N V_i T_i \Delta Z_i - q T_0, \quad (3)$$

where $S_0 = 34.8$ [the nominal value of offshore, upper ocean salinity, following Mertz et al. (1993), and Loder et al. (1998)], $T_0 = 8^\circ C$ are the reference values of salinity and temperature and ΔZ_i ($= 30, 50, 30$ m, $i = 1, 3$ for C2; $= 30, 45, 50, 50, 35$ m, $i = 1, 5$ for NECE) are the representative depth ranges. These functions (q, b, f) are actually *differential* transport estimates, intended to be integrated laterally over relevant portions of a vertical section to provide total transports (see box model below). Missing salinity data were estimated using linear interpolation/extrapolation of data from adjacent instruments with which correlations were typically high ($r^2 \geq 0.85$). The resulting transport functions were then fit with an annual harmonic plus mean (Table 1(b)). Finally, the NECE transport was split into two parts, shallow (< 75 m) and deep (≥ 75 m), to reveal phase differences between the layers and to facilitate comparison with Ramp et al. (1985).

2.2. Hydrographic observations

Hydrographic data from within the Gulf of Maine came from the GLOBEC broadscale survey cruises and from surveys conducted by Northeast Fisheries Science Center (NEFSC) for assessing fish stocks. The GLOBEC broadscale surveys were conducted annually in 1995 (February–July) and 1996–1997 (January–June), sampling Georges Bank on a standard grid of about 40 stations and collecting CTD profiles using a Neil Brown MK5 CTD system with T, S accuracies of $\pm 0.01^\circ C$, ± 0.01 , respectively. On the NEFSC stock assessment cruises, CTD observations were made routinely with a SeaBird SeaCat instrument, either in a standard vertical profile cast or attached to a double oblique plankton tow. In the latter case, the data collected on the in-haul portion of the tow were processed as a vertical profile. The T, S accuracies of the SeaCat in this

mode of operation were roughly $\pm 0.01^\circ\text{C}$, ± 0.01 , respectively. The timing and location of observations on these fisheries surveys varied, but 30 or more stations were occupied on both Georges Bank and in the Gulf of Maine at a rate of roughly six times per year.

For each cruise (GLOBEC or resource survey), the average surface layer temperature and salinity (0–30 m) were calculated in each of four regions representing the three major basins (Wilkinson, Jordan and Georges) and the northwest quarter of Georges Bank (Fig. 1b). The northwest quarter of the Bank was selected to avoid the confounding influences of intrusions of waters directly from the Scotian Shelf and from the Slope Water region to the south.

2.3. Oxygen isotope data

Coincident oxygen isotope and salinity samples were collected at standard depths on all GLOBEC broadscale surveys, Bedford Institute of Oceanography (BIO)/GLOBEC mooring cruises and other selected BIO cruises on the Scotian Shelf and in the Gulf of St. Lawrence (Fig. 1b). Isotope analysis consists of measuring the $\text{H}_2^{18}\text{O}/\text{H}_2^{16}\text{O}$ abundance ratio with a VG PRISM II mass spectrometer using the method of Epstein and Mayeda (1953). The ^{18}O depletion index is expressed in the $\delta^{18}\text{O}$ notation, defined as the per mil fractional difference of the $\text{H}_2^{18}\text{O}/\text{H}_2^{16}\text{O}$ ratio in the sample from that of a standard:

$$\delta^{18}\text{O}(\text{‰}) = \frac{(\text{H}_2^{18}\text{O}/\text{H}_2^{16}\text{O})_{\text{sample}} - (\text{H}_2^{18}\text{O}/\text{H}_2^{16}\text{O})_{\text{std}}}{(\text{H}_2^{18}\text{O}/\text{H}_2^{16}\text{O})_{\text{std}}} \times 1000.$$

The reference used in this study is VSMOW (Vienna Standard Mean Ocean Water, Coplen, 1994). Precision of the mass spectrometer results in a $\delta^{18}\text{O}$ standard deviation of $\sigma = 0.005$. The analysis process involves a gas exchange between CO_2 and H_2O , performed in batches of 60 samples. Each water sample is run in replicate along with 12 standards. Precision of $\delta^{18}\text{O}$ for the standards is $\sigma = 0.033$ over the course of many batches and this can be considered the accuracy of a sample $\delta^{18}\text{O}$. A sample is reanalyzed if the replicate $\delta^{18}\text{O}$ difference is greater than 2σ .

2.4. Freshwater sources

To investigate the particular sources of freshwater contributing to the Gulf of Maine/ Georges Bank system during the GLOBEC Phase I field years, climatological indices of monthly- and annual-mean data were acquired from a number of sources including the BIO environmental data archive (available on the DFO website: <http://www.mari-times.dfo.ca/science/ocean/>), the NOAA data archive (<http://www.ncdc.noaa.gov/pub/data/coop-precip.html>), and the REDIMS distributed data archive for the Gulf of Maine (<http://oracle.er.usgs.gov/GoMaine/histream.html/>). The selected time series, categorized as of local or remote origin, are listed below along with their source archive:

Local

- Boston precipitation (NOAA).
- Maine river discharge (REDIMS).
- St. John River discharge (BIO).

Remote

- RIVSUM (BIO): a composite measure of St. Lawrence river discharges consisting of the monthly averaged flow rates of the St. Lawrence River at the Cornwall dam, the Ottawa River at the Carillon dam and the Saguenay River at the Isle Maligne dam.
- Gulf of St. Lawrence ice cover (BIO).
- Labrador Shelf ice cover south of 55°N (BIO).
- Station 27 salinities at 0 and 50 m (BIO): a standard monitoring station (47°33'N, 52°35'W) roughly 8 km off the mouth of St. John's Harbour on the Newfoundland Shelf.

3. Results

3.1. Boundary fluxes

The means and annual cycles of currents, temperatures and salinities at C2 and NECE (Table 1) reveal some familiar features of the inflow (e.g., C2 surface inflow peaks in early winter (Smith, 1989b); NECE deep inflow peaks in late summer (Fig. 2c; also see Ramp et al., 1985)), and also provide some new insights. A strong surface (< 75 m) inflow at NECE peaks in spring (March–April; Fig. 2a) as opposed to late summer as the deep flow does. This suggests that processes governing shallow and deep inflows may be different. Specifically, the shallow inflow peak appears to coincide with the arrival of the freshwater pulse from the Gulf of St. Lawrence on outer Browns Bank (Smith, 1983), which would affect the surface layers at NECE most strongly. Unfortunately, it is not possible to confirm this effect directly using the annual cycle for near-surface salinity at NECE since the amplitude and phase estimates are not statistically significantly different from zero. In fact, there are no statistically significant annual salinity cycles found at any depth (except near bottom) at the NECE site. Possible explanations for this are (1) that salinity variations are controlled to a large extent by offshore processes (see discussion), and/or (2) that the rapid change of phase of the annual salinity cycle across Northeast Channel (Mountain and Manning, 1994) confounds its detection. Finally, another new result is the discovery of a distinct annual cycle in NECE cross-channel (U) current, favoring “crossover” flow towards Georges Bank in winter (Figs. 2b and d). This too may be related to the arrival of the freshwater pulse from the Gulf of St. Lawrence, although there does not seem to be a direct correlation between “crossovers” and Gulf of St. Lawrence discharge (Bisagni et al., 1996).

As has been noted in earlier work, annual amplitudes (phases) of temperature decrease (increase) with depth, consistent with downward diffusion of surface heat flux (see Smith, 1989b), and a distinct annual salinity cycle is found at all depths on C2.

The annual transport cycles for volume and freshwater at C2 (Fig. 3a and c) reflect the early winter inflow maxima described by Smith (1983, 1989b), whereas the depth-integrated volumetric transport at NECE (Fig. 3b) differs from that of Ramp et al. (1985), peaking in May–June rather than late summer. The full-depth freshwater transport (Fig. 3d) peaks even earlier (April), coincident with the occurrence of the salinity minimum on the adjacent region of Browns Bank (Smith, 1983). Part of this discrepancy may be explained by splitting the transports into two parts: shallow (< 75 m) and deep (≥ 75 m) (Table 1(b); Fig. 4). The peak of the deep volumetric transport (July) is

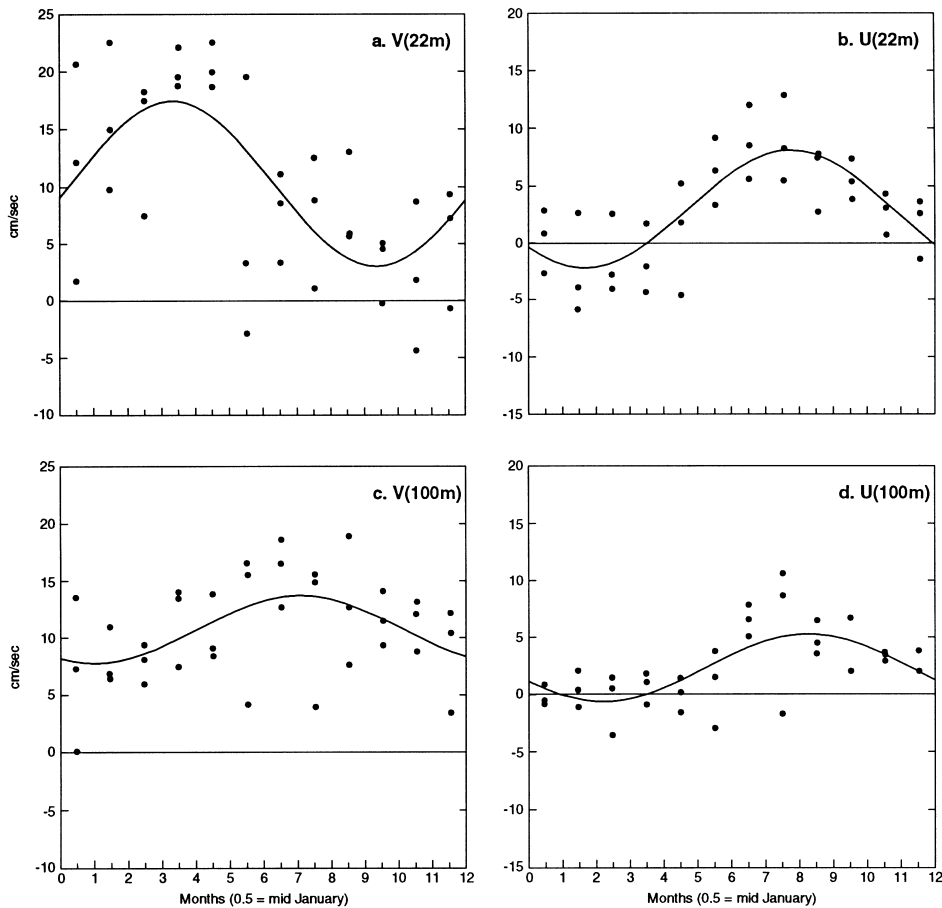


Fig. 2. Monthly mean axial(V) and cross-channel(U) currents from NECE at 22 m (a, b) and 100 m (c,d). Curves indicate harmonic fits to annual cycles plus means.

more consistent with the observations Ramp et al. (1985), whereas the surface layer transport peaks in April–May, as expected from the axial currents (Fig. 2a). The annual cycle of freshwater transport in the shallow layer is reflected in the depth-averaged transport since the deep annual freshwater cycle is insignificant. The annual cycles of heat transport (not shown) basically follow the temperature cycles and are most pronounced in the shallow layers.

The anomaly time series in the NEC deep layer (Fig. 5) reveal low-frequency signals over the 3-yr period that are reasonably coherent over the lower 100 m of the water column, but are obscured by month-to-month variability in the records, particularly in winter. At least some of this high-frequency “noise” may be traced to exceptional wind events (e.g., in February, 1995; R. Limeburner, personal comm.), which have a strong impact on the deep inflow currents and properties (Ramp et al., 1985). Ignoring the high-frequency variability, there appear to be three distinct low-frequency events in the deep NECE records:

1. a period of enhanced warm, salty inflow at the beginning (October 1993–September 1994), followed by

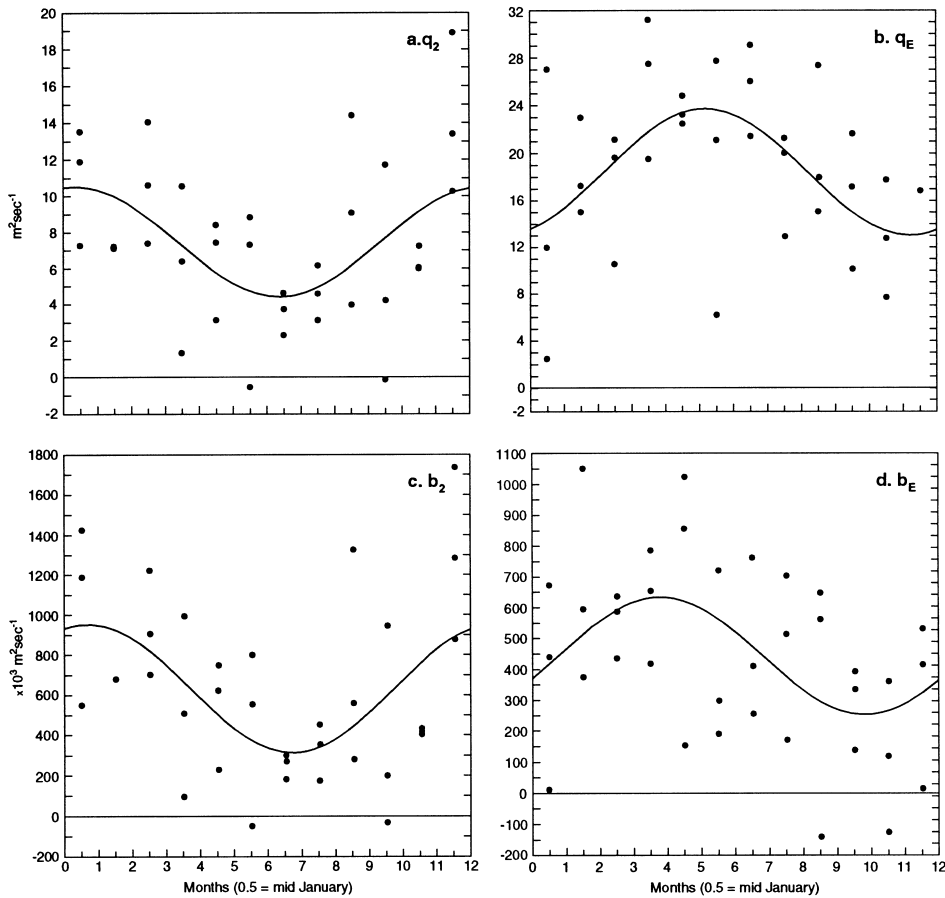


Fig. 3. Monthly mean volumetric, q , and freshwater, b , differential transports at C2 (a, c) and NECE (b,d). Curves indicate harmonic fits to annual cycles plus means.

2. a sharp reduction of the inflow and transition to cooler, fresher than normal conditions during the winter of 1994/1995, and
3. a second period of greatly reduced inflow and cold, fresh conditions during the first nine months of 1996.

The ranges of inflow current (15 cm s^{-1}), temperature (5°C), and salinity (1.5) anomalies at the 100m level are large, particularly those for T and S , which rival those associated with decadal variability (Petrie and Drinkwater, 1993). The T, S properties of the deep inflow (Fig. 5d) indicate that at 100 m, the water is basically a mixture of mid-depth SSW (2°C , 32.0; Houghton et al., 1978) and Gatién's (1976) Warm Slope Water (WSW; 11°C , 35.0), whereas at the deeper levels (150, 190 m), the mixtures include some Labrador Slope Water (LSW; 6.5°C , 34.5) and Gulf Stream (GS, 16°C , 36.0; Houghton and Fairbanks, 2001).

To emphasize the low-frequency signals, a 5-month running mean filter was applied to the transport anomaly data. The events in the deep inflow (Fig. 6c) are now clearly identifiable, with

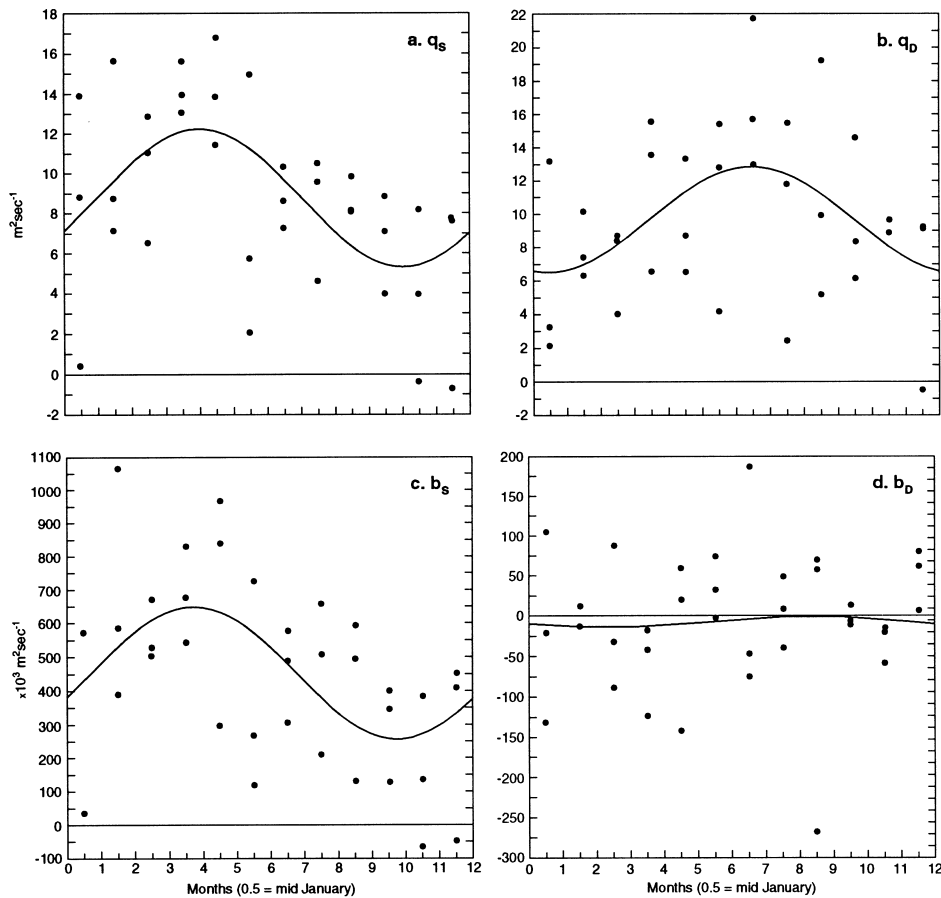


Fig. 4. Monthly mean volumetric, q , and freshwater, b , differential transports at shallow (< 75 m) NECE (a, c) and deep (> 75 m) NECE (b, d). Curves indicate harmonic fits to annual cycles plus means.

peak inflow anomaly occurring in July 1994, followed by the winter 1994/1995 transition, then persistent negative anomalies in 1996. The inflow transport anomalies at C2 (Fig. 6a) are clearly out of phase with those of deep NECE, featuring reduced inflow in the early period and enhanced inflow in late 1994 and in 1996. The character of events is somewhat different in the NECE surface layers (Fig. 6b) where the transition to reduced inflow occurs earlier in 1994, the 1996 reductions are weaker, and 1995 features enhanced inflow.

The freshwater transport anomalies (Fig. 7) generally reflect the volumetric anomalies, occurring in phase in the surface layer and out of phase in the deep NECE layer. This merely signifies that the transport-weighted salinity exceeds the reference salinity ($\bar{S} > S_0$, where $\bar{S} = q^{-1} \int_{-H}^0 V S dz$; Mertz et al., 1993) in the deep water, so that inflow salinates rather than freshens the Gulf. The opposite is true in the surface layers. Due to the relative weakness of the deep NECE freshwater fluxes, the net transport anomalies (not shown) are similar to those in the surface layer, i.e. significantly negative for most of 1994 and strongly positive following a transition period in March–April 1995. Taken together, the C2 and NECE flux anomalies indicate freshening inflow to the Gulf was occurring

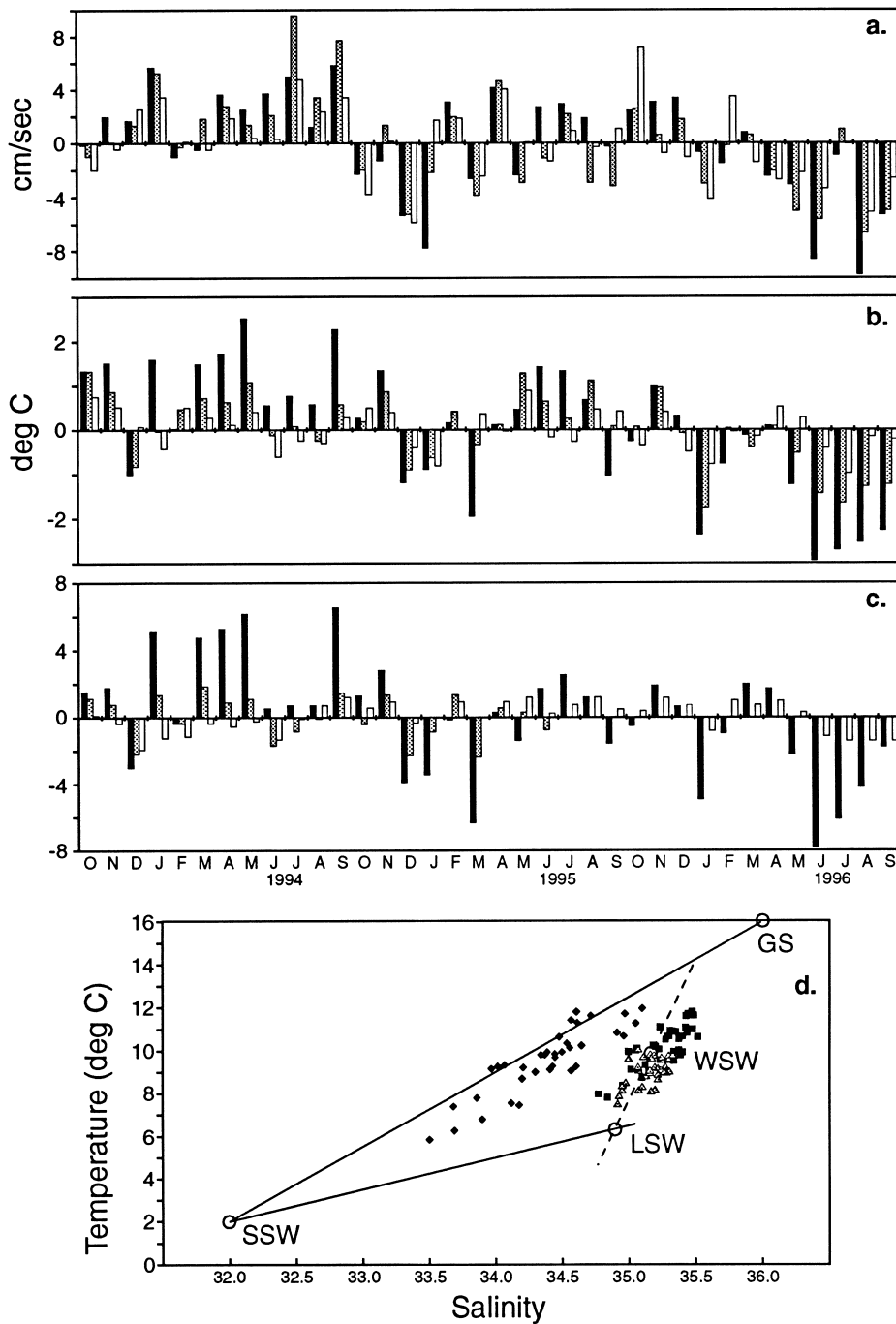


Fig. 5. Anomaly time series in the NECE deep layer [100 m (solid), 150 m (shaded), 190 m (open)] including time series for (a) inflow current, (b) temperature, (c) salinity. Monthly mean temperature vs. salinity data at 100 m (diamonds), 150 m (squares) and 190 m (triangles) are shown in (d). Open circles in (d) indicate nominal properties of Scotian Shelf Water (SSW), Warm Slope Water (WSW), Labrador Slope Water (LSW), and Gulf Stream (GS), according to Houghton and Fairbanks (2001).

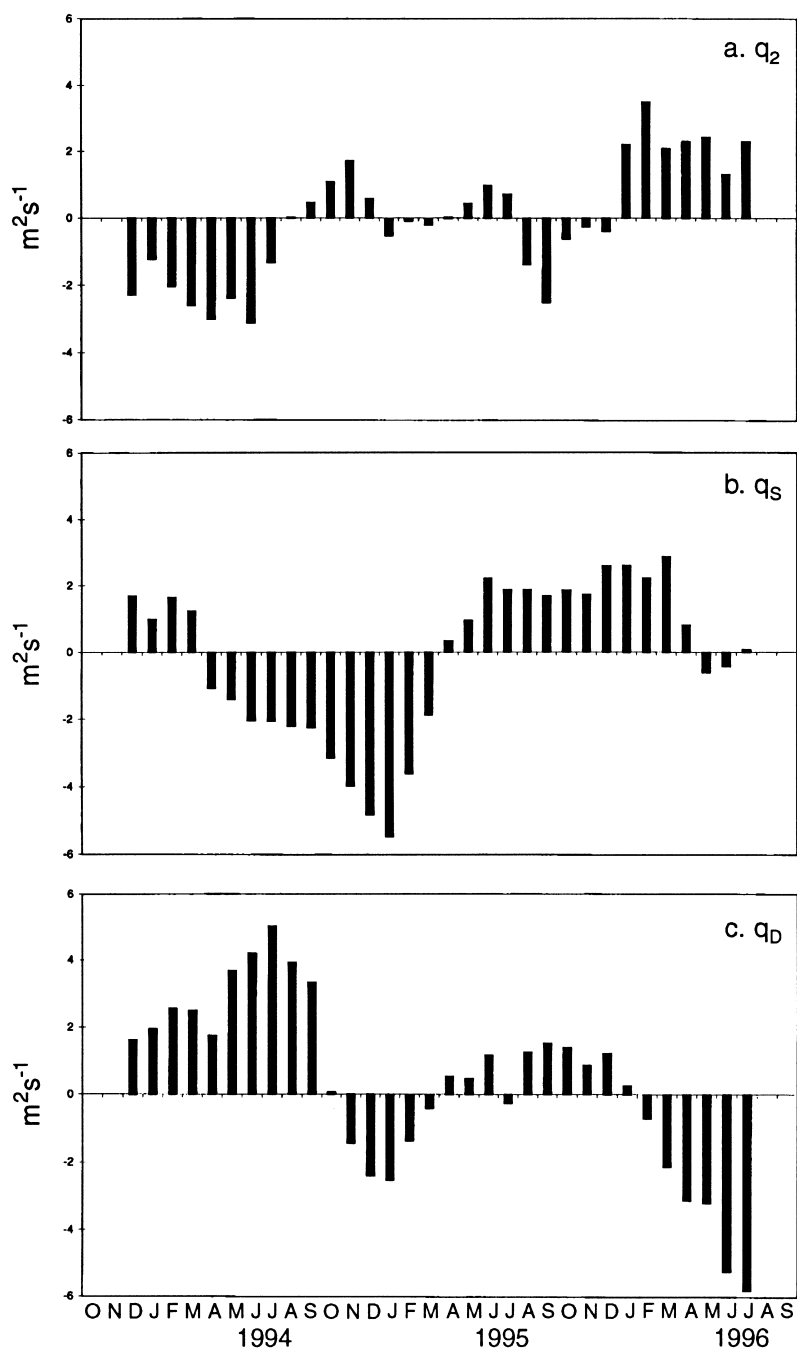


Fig. 6. Time series of differential volumetric transport anomalies at (a) C2, (b) NECE shallow layer (< 75 m), and (c) NECE deep layer (> 75 m). A 5-month running mean has been applied to the monthly data.

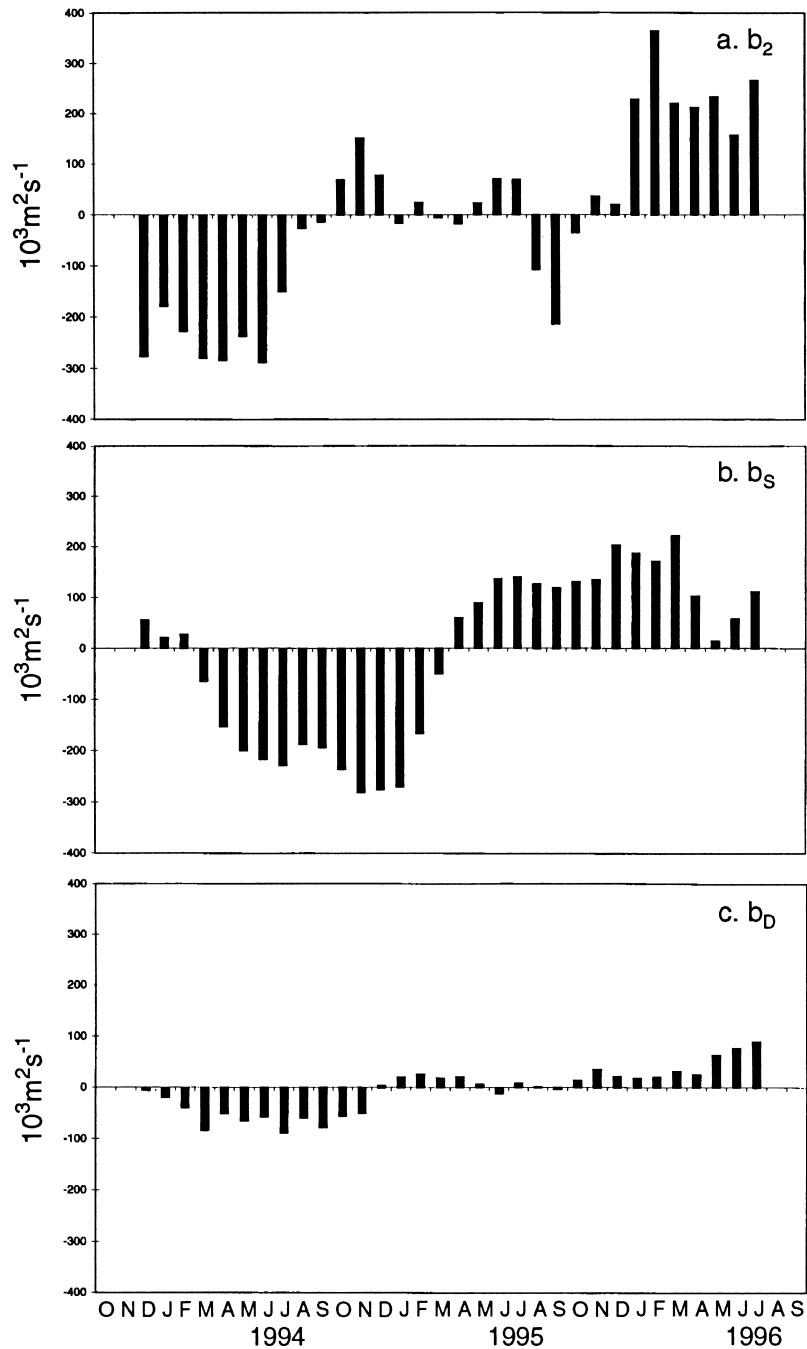


Fig. 7. Time series of differential freshwater transport anomalies at (a) C2, (b) NECE shallow layer (< 75 m), and (c) NECE deep layer (> 75 m). A 5-month running mean has been applied to the monthly data.

from spring 1995 onward, with the strongest fluxes occurring in 1996 to the end of the record. [More recent data (September 1996–May 1997) show a continuation of anomalously high fresh-water inflow in the NECE surface layer.]

3.2. Water properties in the gulf

The NEFSC surface layer (0–30 m) salinity anomalies relative to the MARMAP climatology show behavior consistent with that of the boundary fluxes during 1993–1996 (Fig. 8). Positive anomalies occur only in 1994 in the inner basins, but persist on NW Georges Bank into early 1995. (The isolated positive anomaly on the Bank in late 1995 is associated with an encroachment of the shelf/slope front onto the southern flank, not an increase in the salinity of the Bank water itself.) The increased salinities reflect primarily the reduced freshwater fluxes occurring on the boundaries in late 1993 through early 1995 (Fig. 7; also see box model below). Although the phase of the high-salinity pulse is not well resolved, it does appear to progress counterclockwise around the Gulf, arriving on Georges Bank toward the middle of 1994, 8–9 months after its first detection at C2 or NECE. Preceding the brief period of positive salinities was another protracted period of negative anomalies, such that overall, the surface layers of the Gulf were decidedly fresher during the 1990s than during the MARMAP period.

Similar events were detected in the deep waters of the Gulf (150–200 m, Fig. 9) where salinities declined from 1991 to mid-1993, then rose to the end of 1994, fell again from mid-1995 through early 1996, then recovered. [The long-term trends are partly obscured by the annual deep water cycle that has a minimum in spring and maximum late in the year (Mountain and Manning, 1987).] The recovery to normal salinity levels in early 1997 was also noted in records at 0 and 90 m in the

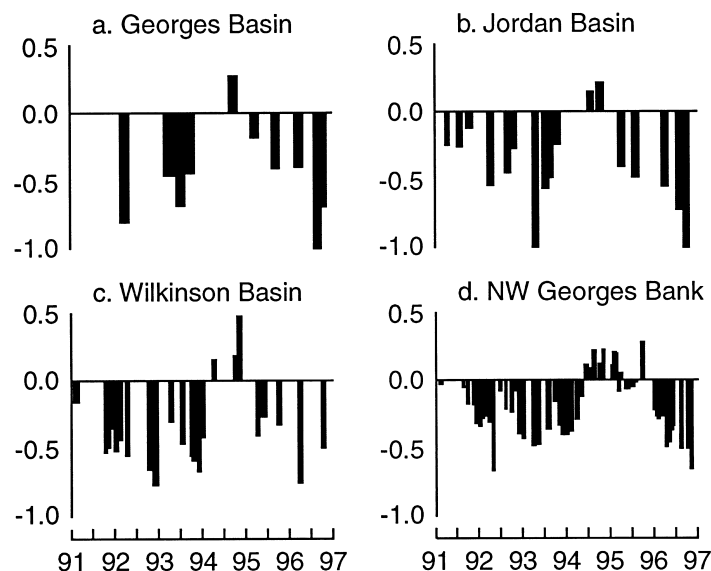


Fig. 8. Surface layer (0–30 m) salinity anomalies for selected areas of the inner Gulf of Maine and Georges Bank (see Fig. 1b). Residuals are calculated with respect to climatological annual cycles for the MARMAP period (1977–1987).

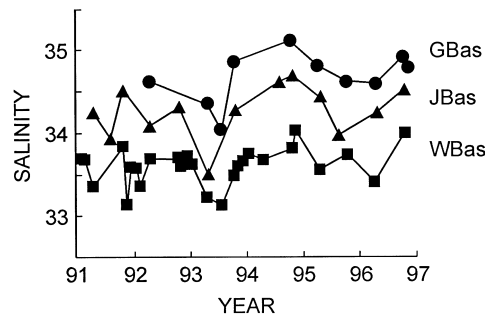


Fig. 9. Deep layer (150–200m) average salinities in the inner basins of the Gulf of Maine: 1991–1997.

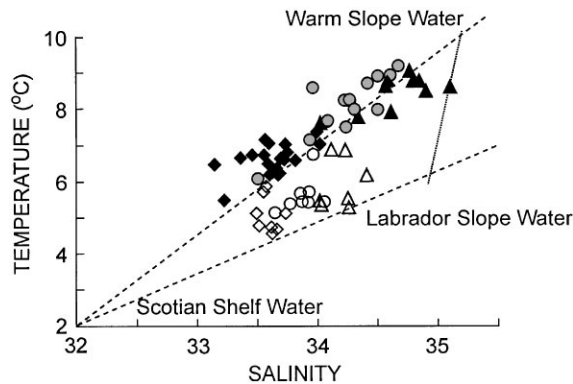


Fig. 10. Temperature vs. salinity for deep basin layers (150–200m) comparing the 1990s (solid) to the 1960s (open). Triangles, circles and diamonds represent Georges, Jordan and Wilkinson Basins, respectively.

Bay of Fundy off St. Andrews, N.B. (Drinkwater et al., 1998). The T, S characteristics of the deep waters (Fig. 10) lie mostly along the same mixing line connecting SSW and WSW, except for some observations in Georges Basin in 1994 that are shifted toward LSW. These waters may originate in the deepest layers of the NEC inflow (Fig. 5d) and remain trapped in Georges Basin by the sills separating it from adjacent basins (e.g., Truxton Swell at the boundary to Jordan Basin). In contrast, conditions from the same areas during the cold 1960s were significantly colder and fresher, indicating the presence of a large component of LSW. Thus it appears that the fluctuations of Gulf water properties during the 1990s is caused by variable mixing ratios between the two end points (SSW, WSW) rather than a change in the character of the slope water end point itself.

The dramatic changes in surface layer salinity on NW Georges Bank were captured by the GLOBEC broadscale surveys in 1995–1997 (Fig. 11a). The lowest average salinities near 32.0, recorded in 1997, were 0.5–1.1 less than those found during the same month in 1995. In contrast, the lowest temperatures, recorded in 1996, differed from those in 1995 by less than 2°C (Fig. 11b), which is small relative to the amplitude of the annual cycle, but may be biologically significant. Nevertheless, the unmistakable conclusion is that a strong interannual pulse of freshwater influenced the water properties in the entire Gulf of Maine during 1995 to mid-1997. Our next task is to investigate its source.

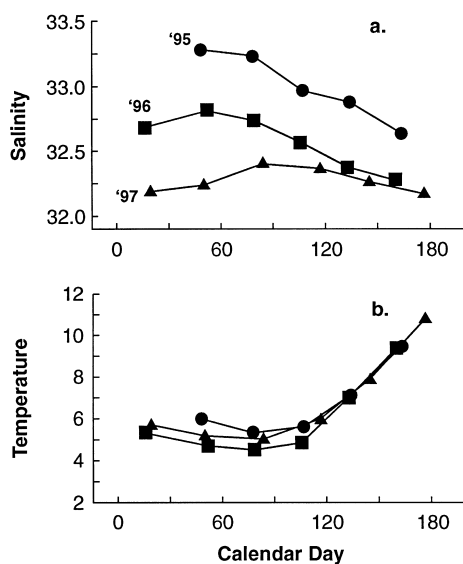


Fig. 11. Monthly average (a) salinity and (b) temperature in the surface layer (0–30 m) on NW Georges Bank (see Fig. 1b) from the GLOBEC broadscale surveys: 1995–1997.

3.3. Oxygen isotope analysis

Water properties of the Gulf of St. Lawrence and Scotian Shelf are derived from mixtures of St. Lawrence River Water (SLRW), Labrador Shelf Water (LShW), deeper Labrador Slope Water (LSW), Warm Slope Water (WSW), and Gulf Stream (GS) water. Warm Slope Water (WSW) and Gulf Stream (GS) derivatives were not observed in the Gulf of St. Lawrence, but first appeared along the Scotian Shelf and in Northeast Channel. The $\delta^{18}\text{O}$ -S values defining these water types (Fig. 12a) are chosen to be the most appropriate end members of the mixing curves observed in the data being considered here. A proxy for SLRW (Fig. 12a) is St. Lawrence Estuary Water (SLEW), which represents brackish SLRW found in the western Gulf of St. Lawrence out to the western side of Cabot Strait. Likewise, MCW (Maine Coastal Water) is a proxy for Maine River outflow that mixes into the interior of the Gulf of Maine at roughly 150 m depth. At the western end of the Scotian Shelf, $\delta^{18}\text{O}$ vs. S data for the inflow waters off Cape Sable and in the NEC in 1995 (June, November) are tightly distributed along a single mixing curve (the individual regressions are statistically indistinguishable) for which the freshwater end member itself lies on a mixing line between SLEW and LShW (Fig. 12a). By volume, SLRW contributes only 5.6% of this mixture, but represents 48% of the freshwater content (relative to $S_0 = 34.8$). This compares favorably with the 55% estimated by Koutitonsky and Bugden (1991) and Loder et al. (1998). Subsequent data in September 1996 and July 1997 define the same mixing curve, so properties of the water entering the Gulf of Maine via NEC and Cape Sable appear to be invariant during the years 1995–1997.

Studies by Loder et al. (1998) estimate a climatological mean freshwater flux (relative to 34.8) through the NEC of $13 \times 10^3 \text{ m}^3 \text{ s}^{-1}$ and a river runoff into the Gulf of Maine of $3 \times 10^3 \text{ m}^3 \text{ s}^{-1}$. Thus Maine rivers contribute approximately 18% of the overall freshwater input into the Gulf of

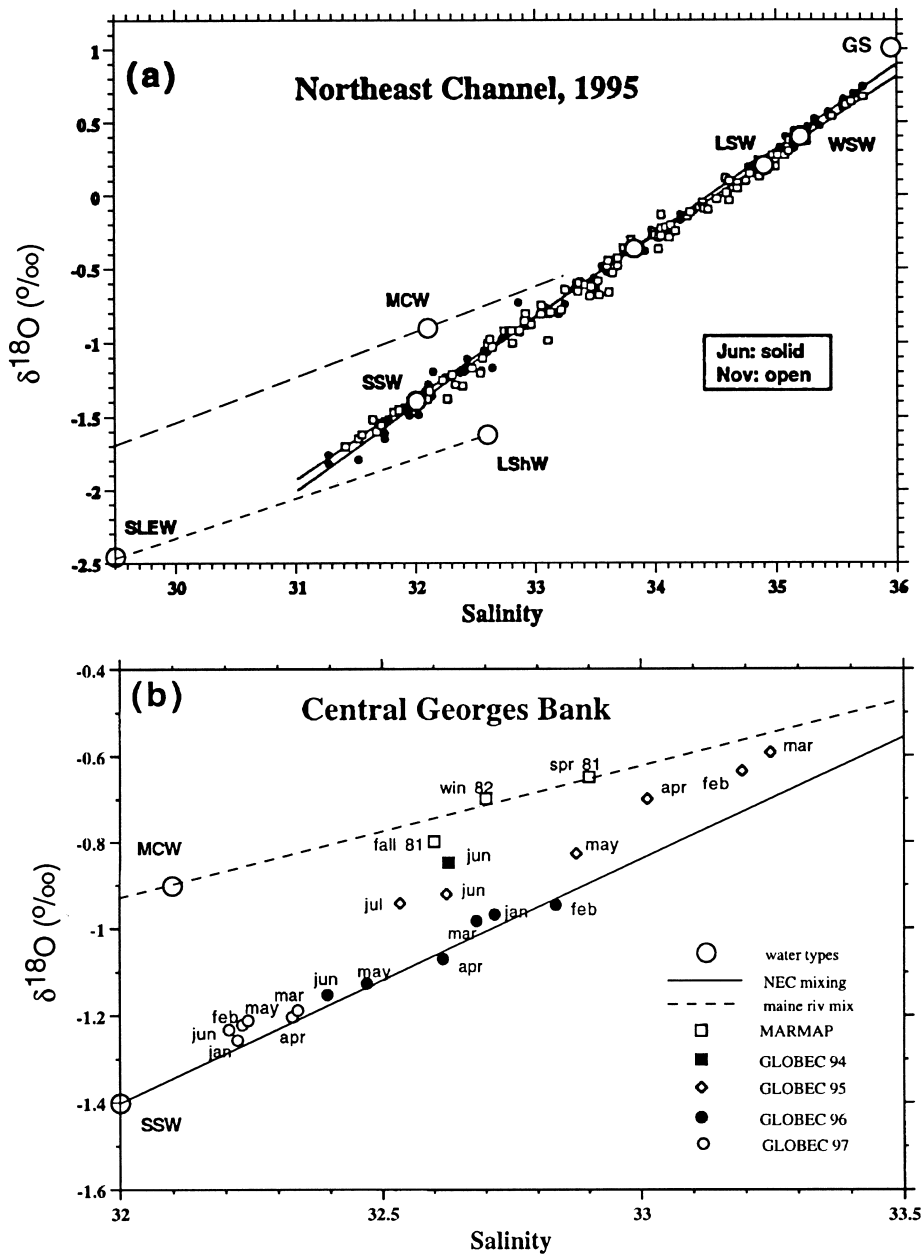


Fig. 12. Oxygen isotope depletion index ($\delta^{18}\text{O}$) vs. salinity (a) for 1995 samples (June and November) from the western Scotian Shelf and Northeast Channel, and (b) for central Georges Bank from the GLOBEC broadscale surveys: 1994–1997 and MARMAP: 1981–1982. Open circles represent canonical values for shelf and slope water masses: St. Lawrence Estuary Water (SLEW), Labrador Shelf Water (LShW), Labrador Slope Water (LSW), Warm Slope Water (WSW) and Gulf Stream (GS) are defined as appropriate end members for mixing curves derived from 1995 data. The value for Scotian Shelf Water (SSW) is an approximate mean value; it fluctuates seasonally (fresher in February) along the regression line. The Maine Coastal Water (MCW) value represents a proxy for Maine River water that mixes into the interior of the Gulf of Maine. Solid lines represent linear regressions in western SS/NEC for (a) 1995 and (b) a 1995–1997 composite. Dashed and dotted lines are mixing curves for St. Lawrence and Maine River waters respectively.

Maine. Much of this freshwater may be discharged in the cyclonic coastal current that exits the Gulf over Nantucket Shoals and does not necessarily mix thoroughly throughout the interior of the Gulf.

On central Georges Bank, the monthly series of oxygen isotope properties from the GLOBEC broadscale surveys (Fig. 12b) shows a persistent trend between 1994 and 1997, away from mixtures of local Maine Coastal Water (MCW) and toward the NEC–Scotian Shelf mixing line. Thus the significant freshening of the Bank, earlier noted in Fig. 11a, also appears to represent the exclusion of waters from Maine rivers. The dominant seasonal variation during 1996–1997 is due to changes in mixing ratios along this curve, with Maine rivers contributing less than 5% of the freshwater on the Bank. In 1995, the Georges Bank water (GBW) appears to lie between the SSW and the MCW mixing lines, along a curve whose freshwater end member represents roughly 50% MCW and 50% SSW. On this curve, Maine river water contributes 1.4% by volume and 26% of the freshwater relative to 34.8. In 1994, the coastal influence was even greater, with MCW contributing 63% to the GBW end member and Maine rivers providing 2.4% by volume and 38% of the freshwater. For comparison, data from the MARMAP period (Fig. 12b) suggest that very little SSW was present on Georges Bank during 1981–1982, especially in winter/spring, and that MCW contributed approximately 75% of the freshwater on the Bank.

In summary, the strong fluxes detected on the inflow boundaries of the Gulf of Maine in late 1995 and 1996 appear to have contributed significant amounts of freshwater of remote origin to the Gulf, and especially to Georges Bank during 1996 and early 1997.

4. Discussion

4.1. *Effects of warm core rings*

As has been noticed in previous studies, the presence of warm core rings off the mouth of Northeast Channel is often associated with the presence of Warm Slope Water and/or ring “streamers” (modified Gulf Stream water) in the Channel (Brooks, 1987) and warm, saline conditions off Cape Sable (Smith, 1989b). Similar conclusions apply to the protracted period of enhanced deep NEC inflow that occurred during the first year of moored measurements when conditions were anomalously warm and saline in both the deep layer at NECE (Fig. 5) and at C2 (not shown). Satellite imagery indicates that during this period, the slope water region was populated by a number of rings that intermittently passed near the mouth of the NEC (Fig. 13). On at least one occasion (Fig. 13a), a contemporaneous hydrographic section across NEC at the mooring site reveals the presence of a ring “streamer” with temperatures and salinities in excess of 15°C and 35.5, respectively (Fig. 14). Driven by the enhanced inflow, these properties would then be transported throughout the Gulf by the classic deep-layer circulation described by Brooks (1985) and mixed vertically into the surface layers as suggested by the hydrographic measurements (e.g., Fig. 8). However, the mere presence of a warm core ring near the mouth of NEC is not a reliable indicator of enhanced deep inflow of slope water (Ramp et al., 1985), so that quantitative use of satellite imagery for monitoring major inflow events is not possible.

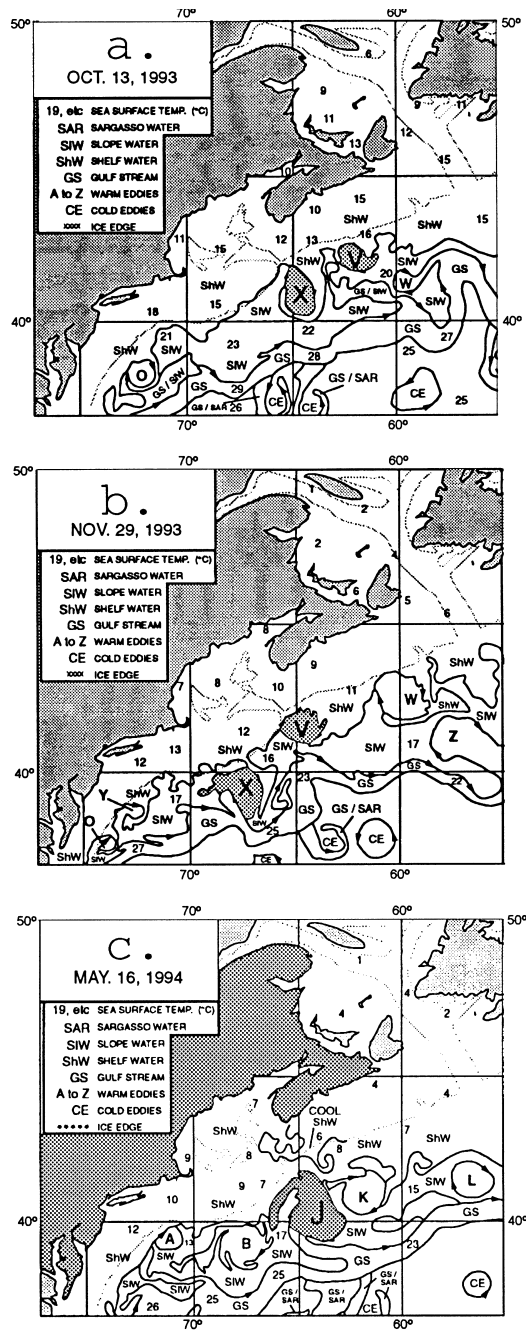


Fig. 13. Surface thermal features map derived from NOAA/NOS/OPC product for (a) October 13, 1993; (b) November 29, 1993, and (c) May 16, 1994. Warm-core rings X, V, and J are highlighted in the vicinity of the mouth of Northeast Channel.

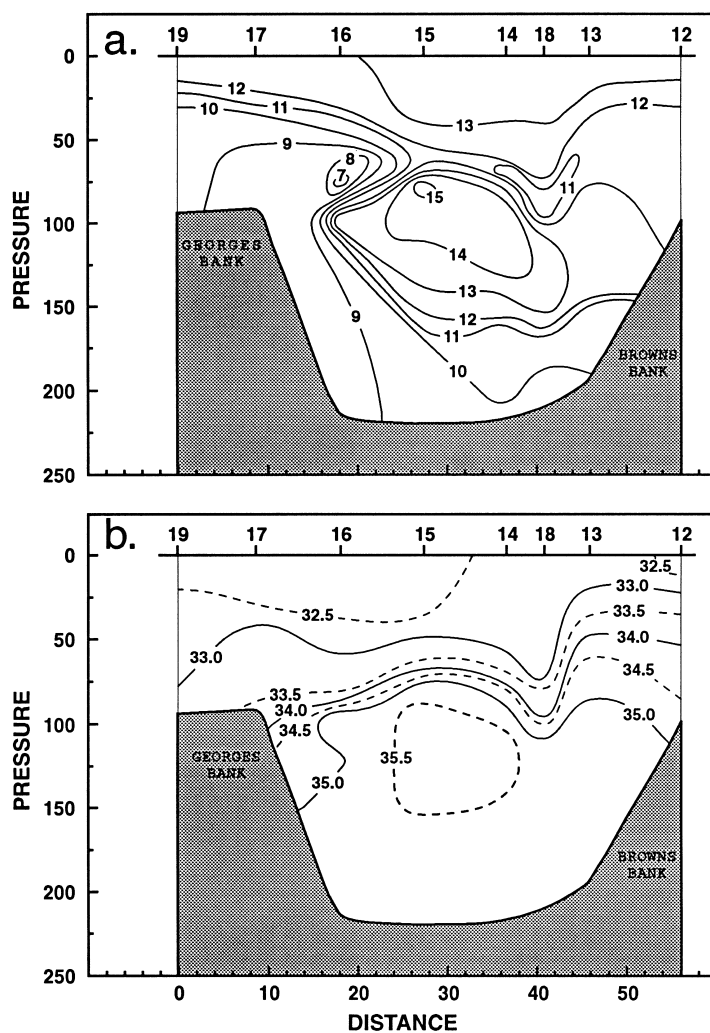


Fig. 14. Hydrographic sections across Northeast Channel at the location of the mooring sites (NECE, NECW; see Fig. 1a) conducted during 14–15 October, 1993.

4.2. Boundary fluxes

The boundary flux measurements at C2 and NECE provide crude indices for the net transports into the Gulf from the Scotian Shelf and in the deep layer of NEC. The choice of the C2 mooring site is based on more comprehensive transport estimates from an array of four moorings stretching from the 60 m isobath off Cape Sable to the 100 m isobath on the offshore edge of Browns Bank (C1–C4; Smith, 1983). These measurements indicated: (1) a strong coherence among the inflow currents at low frequencies, and (2) a concentration of the net Scotian Shelf Water transport in the inshore zone (near C1 and C2) because of the presence of a clockwise gyre over Browns Bank (C3 and C4). Based on these findings, it is possible to derive a linear regression for the fraction of

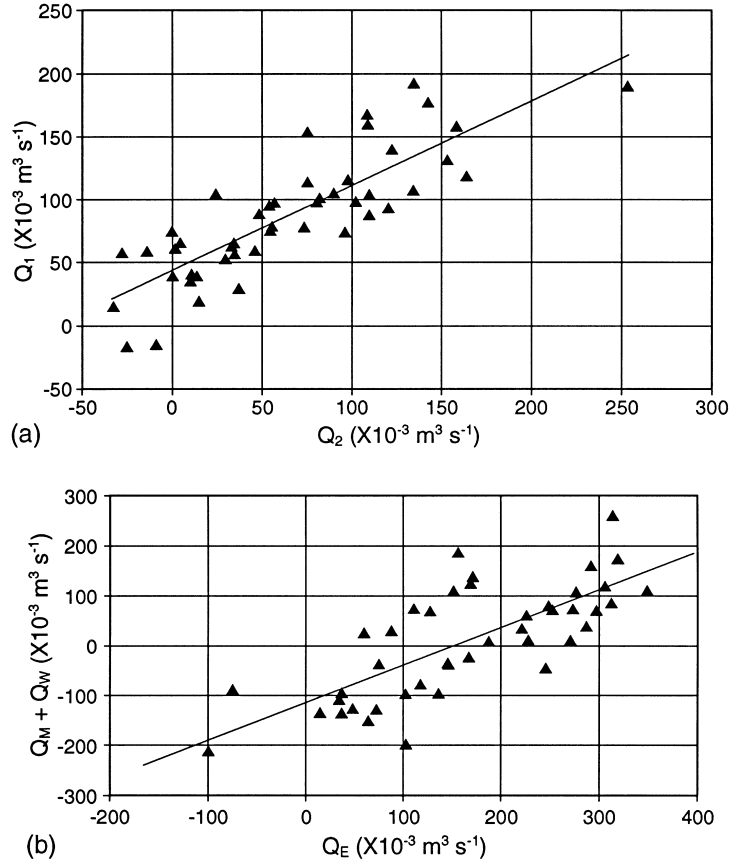


Fig. 15. Relationships among components of estimated monthly volumetric transport (a) off Cape Sable, based on earlier measurements by Smith (1983), and (b) in deep Northeast Channel, based on earlier measurements by Ramp et al. (1985). The inshore component at Cape Sable (Q_1) is related to the C2 observed value (Q_2); the deep NEC transports in the middle (Q_M) and western (Q_W) sides of the Channel are related to those at NECE on the east (Q_E). Regression estimates for the extra components of the transports are

$$Q_1 = 0.68 Q_2 + 45 \times 10^3 \text{ m}^3 \text{ s}^{-1} (r^2 = 0.69),$$

$$Q_M + Q_W = 0.75 Q_E + 112 \times 10^3 \text{ m}^3 \text{ s}^{-1} (r^2 = 0.55).$$

volumetric transport in the nearshore zone (centered at C1 on the 60 m isobath) upon the volumetric transport centered on C2 (Fig. 15a), and then to construct an estimate of the total net transport past Cape Sable, i.e.

$$Q_{CS}(\text{m}^3 \text{ s}^{-1}) = Q_1 + Q_2 = 1.68 Q_2 + 45 \times 10^3 = 1.68 q_2 W_2 + 45 \times 10^3, \quad (4)$$

where (Q_1, Q_2) are the components of the volumetric transport centered on the (C1, C2) mooring sites and W_2 ($= 20 \text{ km}$) is the effective width of the Cape Sable section over which the differential transport function, q_2 , applies. This regression suggests that the transport is distributed rather uniformly over the nearshore-zone.

Similarly, the biweekly NEC transport estimates of Ramp et al. (1985), derived with a considerable amount of interpolation/extrapolation of the measured inflow currents at moorings on the east (E), middle (M) and west (W) sides of the Channel, may be used to relate the net volumetric inflow in the deep NEC layer (> 75 m) to that at the NECE mooring (Fig. 15b), assuming the latter is equivalent to Ramp's E mooring, i.e.

$$Q_D(\text{m}^3 \text{s}^{-1}) = Q_E + Q_M + Q_W = 1.75Q_E - 112 \times 10^3 = 1.75q_D W_E - 112 \times 10^3 \quad (5)$$

where $W_E (= 15 \text{ km})$ is the effective width of the NEC over which q_D applies.

To date, no effort has been made to relate the shallow NECE flux to the net surface inflow in the Channel (due to a lack of data in the near-surface current fields), but its contribution would alter the magnitude and timing of the net NEC inflow significantly, as suggested by the differential transport estimates. The extent to which these NEC surface waters are advected eastward by the Browns Bank gyre and fed into the near-shore flow past Cape Sable (as suggested by drift measurements; Smith, 1989a) is also unclear. However, because of the differences in phase of the annual cycle and anomaly patterns, the two transports are assumed to be independent for our purposes. For lack of any evidence to the contrary, the deep NECE regression (5) will be applied to the surface layers as well in order to estimate the total fluxes.

4.3. A box model

To estimate the impact of the interannual variation in boundary fluxes on the interior properties of the GOM, consider a slightly revised version of Brown and Beardsley's (1978) box model (Fig. 16). In this case, there are five sources of volume and salt flux into the box:

1. Q_{CS} and F_{CS} are the volume and salt fluxes from the Scotian Shelf past Cape Sable, defined in terms of the differential fluxes as

$$Q = \iint V \, dy \, dz \approx W \int_{-H}^0 V \, dz = qW, \quad (6)$$

$$F = \iint VS \, dy \, dz \approx W \int_{-H}^0 VS \, dz = S_0(Q - B) = S_0 W(q - b), \quad (7a)$$

where

$$B = \frac{1}{S_0} \iint (S_0 - S) V \, dy \, dz \approx \frac{W}{S_0} \int_{-H}^0 (S_0 - S) V \, dz = bW, \quad (7b)$$

where S_0 is the reference salinity and W is the width of the inflow current.

2. Q_S and F_S are the shallow (< 75 m) volume and salt influxes in Northeast Channel,
3. Q_D and F_D are the deep (> 75 m) volume and salt influxes in Northeast Channel,
4. Q_R is the river runoff into the Gulf, and
5. Q_{P-E} is the difference between precipitation and evaporation.

The two sinks of mass and salt for the model are the longshore, Q_N and $F_N (= S_N Q_N)$, and offshore, αQ_N and αF_N , fluxes, which are assumed to remain in a fixed ratio and have the same transport-weighted salinity, S_N .

Fig. 16. Box model for volumetric and freshwater transports in the Gulf of Maine. Input transports (Q_{CS}, F_{CS}), (Q_S, F_S), and (Q_D, F_D) enter via Cape Sable, shallow and deep NEC, respectively, through sections of width, W_{CS} and W_{NEC} . River discharge (Q_R) and precipitation–evaporation (Q_{P-E}) enter from land and surface boundaries, and (Q_N, F_N) and $\alpha(Q_N, F_N)$ exit the Gulf via the New England Shelf and the offshore, respectively. A and H are the estimated surface area and mean depth of the Gulf.

Conservation of mass and salt for this system dictate

$$Q_{\text{CS}} + Q_{\text{S}} + Q_{\text{D}} + Q_{\text{R}} + Q_{\text{P-E}} - (1 + \alpha)Q_{\text{N}} = 0 \quad (8)$$

and

$$F_{\text{CS}} + F_{\text{S}} + F_{\text{D}} - (1 + \alpha)F_{\text{N}} = 0 \quad (9a)$$

or

$$S_0(B_{\text{CS}} + B_{\text{S}} + B_{\text{D}} + Q_{\text{R}} + Q_{\text{P-E}}) - (1 + \alpha)(S_0 - S_{\text{N}})Q_{\text{N}} = 0. \quad (9b)$$

Now consider two equilibrium states of this system representing the first (early) and last (late) halves of the 3-yr mooring period: October 1993–September 1996. Defining (dS, dQ, dB) as the differences (late–early) in the average salinity, volume and freshwater fluxes between the two states, subtraction of the balance equations yields

$$(1 + \alpha)dQ_N = dQ_{CS} + dQ_S + dQ_D \quad (10)$$

and

$$(1 + \alpha)Q_N dS_N = (1 + \alpha)(S_0 - S_N)dQ_N - S_0(dB_{CS} + dB_S + dB_D), \quad (11)$$

where salinity changes are assumed small relative to the reference and the runoff and net $P-E$ are assumed invariant. Using the linear regressions (4), (5) from the previous subsection gives

$$dQ_{\text{CS}} = 1.68 dq_2 W_2,$$

and

$$dQ_D = 1.75 dq_D W_E,$$

where $(W_2, W_E) = (20, 15)$ km and the changes,

$$dq_2 = 2.32 \text{ m}^2 \text{ s}^{-1}$$

and

$$dq_D = -2.38 \text{ m}^2 \text{ s}^{-1}$$

are calculated as the differences between the average transport anomalies (Fig. 6a and c) over the late and early halves of the records. Again, for simplicity, the change in the surface layer transport in Northeast Channel was computed in the same way as for the deep transport, i.e.

$$dQ_S = 1.75 dq_S W_E,$$

with

$$dq_S = 3.20 \text{ m}^2 \text{ s}^{-1} \quad (\text{Fig. 6b}).$$

Thus, from (10), the net increase in volumetric transport through the system over the latter half of the mooring period is

$$(1 + \alpha)dQ_N = 1.68dq_2 W_2 + 1.75(dq_S + dq_D)W_E = 99.5 \times 10^3 \text{ m}^3 \text{ s}^{-1}. \quad (12)$$

By comparison, the major components of the total transport through the Gulf under average conditions may be obtained from the mean differential values and the regressions (4), (5); (again assuming equivalence of deep and shallow transport relations), i.e.

$$Q_{CS} = 1.68\bar{q}_2 W_2 + 45 \times 10^3 = 297 \times 10^3 \text{ m}^3 \text{ s}^{-1},$$

$$Q_D = 1.75\bar{q}_D W_E + 112 \times 10^3 = 143 \times 10^3 \text{ m}^3 \text{ s}^{-1},$$

$$Q_S = 1.75\bar{q}_S W_E + 112 \times 10^3 = 119 \times 10^3 \text{ m}^3 \text{ s}^{-1},$$

where $\bar{q}_2, \bar{q}_S, \bar{q}_D = (7.5, 8.8, 9.7) \text{ m}^2 \text{ s}^{-1}$ (Table 1(b)). Using Brown and Beardsley's values for the other components of the mass balance, $(Q_R, Q_{P-E}) = (3, 1) \times 10^3 \text{ m}^3 \text{ s}^{-1}$, gives an estimate of the total volumetric flow rate through the Gulf,

$$(1 + \alpha)\bar{Q}_N = 563 \times 10^3 \text{ m}^3 \text{ s}^{-1}. \quad (13)$$

Note that among the dominant components of this balance, the estimated Cape Sable transport is roughly twice that calculated by Smith (1983) for the 1979–1980 period ($140 \times 10^3 \text{ m}^3 \text{ s}^{-1}$), whereas the deep NEC transport is slightly more than half that found by Ramp et al. (1985) for the 1977–1979 period ($260 \times 10^3 \text{ m}^3 \text{ s}^{-1}$). Thus it appears that, compared to previous estimates, the boundary mass flux of Scotian Shelf water off Cape Sable was higher, and the deep NEC inflow was lower during the GLOBEC field years. Nevertheless, the average total volumetric inflow is very close to the estimate of the climatological mean value of Loder et al. (1998) ($500 \times 10^3 \text{ m}^3 \text{ s}^{-1}$). Also note that the flushing time scale, τ , for the GOM implied by (13) is the order of one year, i.e.

$$\tau = \frac{AH}{(1 + \alpha)\bar{Q}_N} \approx \frac{1.6 \times 10^{13} \text{ m}^3}{0.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}} = 3.2 \times 10^7 \text{ s},$$

where A ($\approx 1.6 \times 10^5 \text{ km}^2$) and H ($\approx 10^2 \text{ m}$) are the estimated surface area and average depth of the Gulf.

Finally, the salt balance equations provide an estimate of the change in the average salinity of the outflow waters from the Gulf. According to (11), this change is related to the changes in both the net transport and the individual buoyancy fluxes. Assuming the differential freshwater fluxes may be inflated by the same factors as the volumetric fluxes to produce section-mean transports across the Cape Sable line and NEC gives

$$dB_{CS} = 1.68db_2 W_2 = 8.80 \times 10^3 \text{ m}^3 \text{ s}^{-1}, \quad (14a)$$

$$dB_S = 1.75db_S W_E = 7.43 \times 10^3 \text{ m}^3 \text{ s}^{-1}, \quad (14b)$$

$$dB_D = 1.75db_D W_E = 1.58 \times 10^3 \text{ m}^3 \text{ s}^{-1}, \quad (14c)$$

where the differences in the depth-averaged flux anomalies (Fig. 7a–c) are estimated as, $(db_2, db_S, db_D) = (0.262, 0.283, 0.060) \text{ m}^2 \text{ s}^{-1}$,

Thus the change in average salinity between the late and early portions of the 3-yr mooring period is given by

$$\begin{aligned} dS_N &= (1 + \alpha)^{-1} Q_N^{-1} \{ (1 + \alpha)(S_0 - S_N) dQ_N - S_0 (dB_{CS} + dB_S + dB_D) \} \\ &= (563 \times 10^3)^{-1} \{ 209 \times 10^3 - 620 \times 10^3 \} = -0.73, \end{aligned} \quad (15)$$

where Brown and Beardsley's average outflow salinity ($S_N = 32.7$) has been used. This single representative value is remarkably consistent with the observations of change in both the shallow (Figs. 8 and 11) and deep (Fig. 9) layers of the Gulf over the 3-yr period. More quantitatively, Houghton and Fairbanks (2000) derive from the GLOBEC broadscale surveys an interannual change in the 6-month average Georges Bank salinity of 0.70. Also note that the increase in volumetric flow rate through the Gulf tends to counteract the increases in freshwater flux, which are concentrated in the surface layers. Finally, the net *change* in the freshwater flux to the Gulf (14a)–(14c) over the latter half of the record exceeds the total climatological-mean freshwater flux ($13 \times 10^3 \text{ m}^3 \text{ s}^{-1}$) estimated by Loder et al. (1998).

4.4. Freshwater sources

Time series of freshwater climatological indices, expressed as annual-average anomalies of variables related to potential sources for the Gulf of Maine, indicate that, relative to the MARMAP period (1977–1987), the early 1990s show no evidence for enhanced local sources of freshwater in either runoff or precipitation (Fig. 17a–c). Weakly positive annual discharge anomalies from the St. John River system in the 1990s are clearly exceeded by positive anomalies during most of the MARMAP period, and the discharge anomalies from the US rivers are mostly negative, with the exception of a moderate peak in 1990. The Boston precipitation anomaly is also weak and variable in the 1990s. The net annual precipitation over the GOM is estimated at roughly 10 cm/year, which is less than 5% of the estimated net advective input. Similarly, the index for Gulf of St. Lawrence river discharge, RIVSUM (Fig. 17d), shows enhanced runoff during the MARMAP surveys.

However, indices for ice cover on both the Labrador/Newfoundland Shelf and in the Gulf of St. Lawrence (Fig. 17e and f) indicate positive anomalies for these sources in the early 1990s, with levels

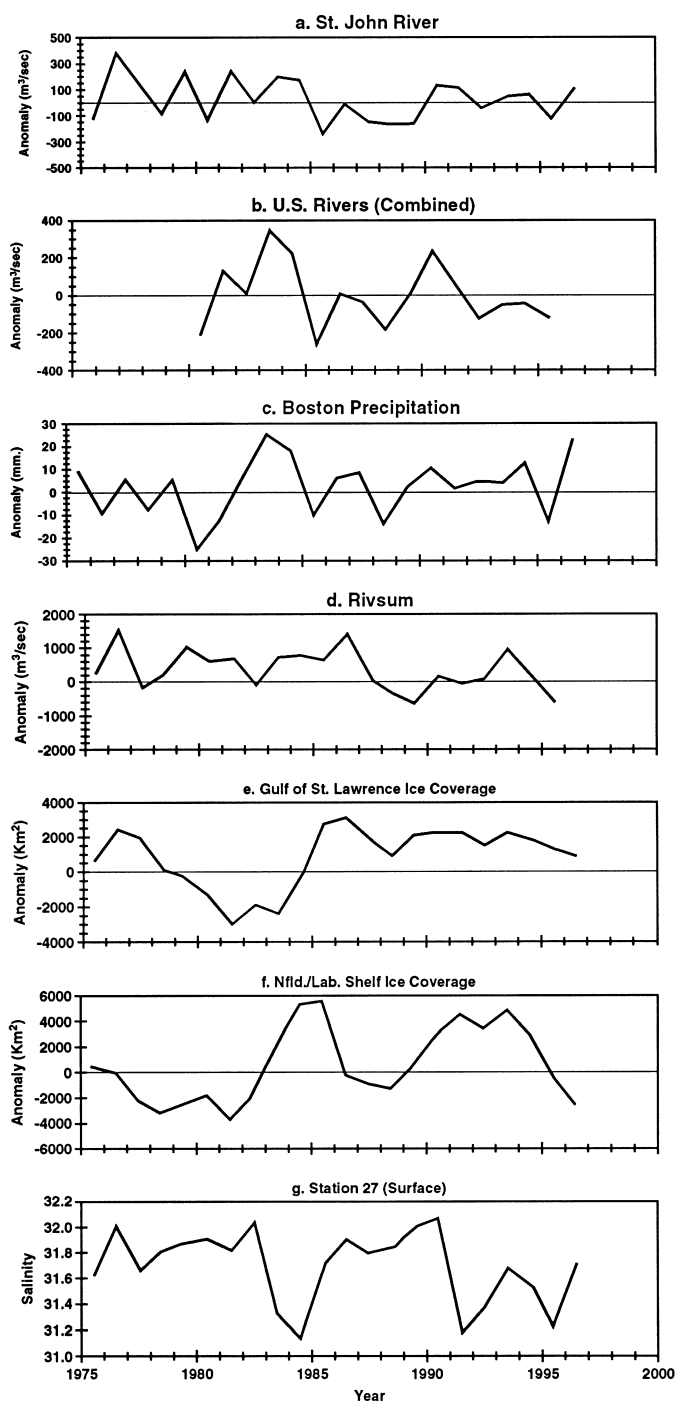


Fig. 17. Time series plots of annual climatological indices for *Local freshwater sources*: (a) Saint John River discharge, (b) Maine river peak discharge (no monthly data available), (c) Boston precipitation, and *Remote freshwater sources*: (d) St. Lawrence River discharge (RIVSUM), (e) Gulf of St. Lawrence ice cover, (f) Labrador/Newfoundland Shelf ice cover south of 55°N , and (g) Station 27 surface (0 m) salinity.

generally in excess of those during MARMAP (except for the period 1983–1986). The much higher ice areas on the Labrador Shelf (vs. in the Gulf of St. Lawrence) also suggest that Labrador is the most likely source of anomalous freshening. Low salinities in the inshore branch of the Labrador Current during the 1990s, detected in surface layer hydrographic measurements at Station 27 (Fig. 17g), are also consistent with enhanced freshwater sources of northern origin. During the MARMAP period, only 1984–1985 showed similarly low salinities at Station 27. Thus it appears that remote freshwater sources in the Labrador Sea or beyond are responsible for the observed freshening of the Gulf of Maine during the mid-1990s.

In an attempt to trace the origin and fate of the freshwater from the Labrador Shelf, irregular monthly time series of upstream near-surface salinity and (negative) sea-ice anomalies are compared to those at the boundary mooring sites and within the Gulf of Maine to NW Georges Bank (Fig. 18). The generally positive anomalies at C2 and NECE during the first year of records (October 1993–September 1994) appear to be related to positive trends in the Bay of Fundy (Prince 5) and on Georges Bank, with phase lags that imply mean advective speeds of $7\text{--}9\text{ cm s}^{-1}$ (i.e. $\approx 3\text{--}4$ months lag between the inflow boundaries and NW Georges Bank). Similar rates may be inferred for the advection of negative anomalies later in the records. These speeds are consistent with those found earlier in MARMAP hydrographic data (Mountain and Manning, 1994).

The structure of negative salinity anomalies at Station 27 off St. John's, Newfoundland, also bears some resemblance to the initial pulses of freshwater at the boundary moorings and is reflected in the scattered observations on the Scotian Shelf. Estimates of advection time ($\approx 3\text{--}4$ months) along the most direct route to the mouth of Northeast Channel ($\approx 1200\text{ km}$ via Haddock Channel, on the southern Newfoundland Shelf, and eastward along the shelf break) give speeds of $12\text{--}16\text{ cm s}^{-1}$, which agree with other observation-based estimates along this trajectory (Marsh et al., 1999). However, subsequent to the freshening at Station 27 in 1995, the anomalies are positive or weakly negative, whereas the anomalies in the Gulf of Maine continue to decline, reaching their respective minima in late 1996 or early 1997 (not shown). For the broad Station 27 salinity minimum in 1995 to have had such an impact, it must have taken a much more circuitous route, e.g., via the Gulf of St. Lawrence. The late freshening is apparent on the inshore Scotian Shelf off Halifax (Fig. 18c) and in the western Cabot Strait (not shown), so a northern source is still implicated.

5. Summary and conclusions

Measurements of the physical properties of the Gulf of Maine/Georges Bank system during the GLOBEC field years 1993–1997 reveal several new features of the annual cycles of the boundary inflows including:

- (a) a strong surface inflow at NECE which peaks in spring (March–April), suggesting processes governing shallow and deep inflows may be different,
- (b) a distinct annual cycle in NECE cross-channel current favoring “crossover” flow towards Georges Bank in winter, and
- (c) the absence of a significant seasonal salinity cycle at any depth at NECE, suggesting that salinity is controlled by offshore processes or that estimation of the cycle is confounded by rapid phase changes in the area.

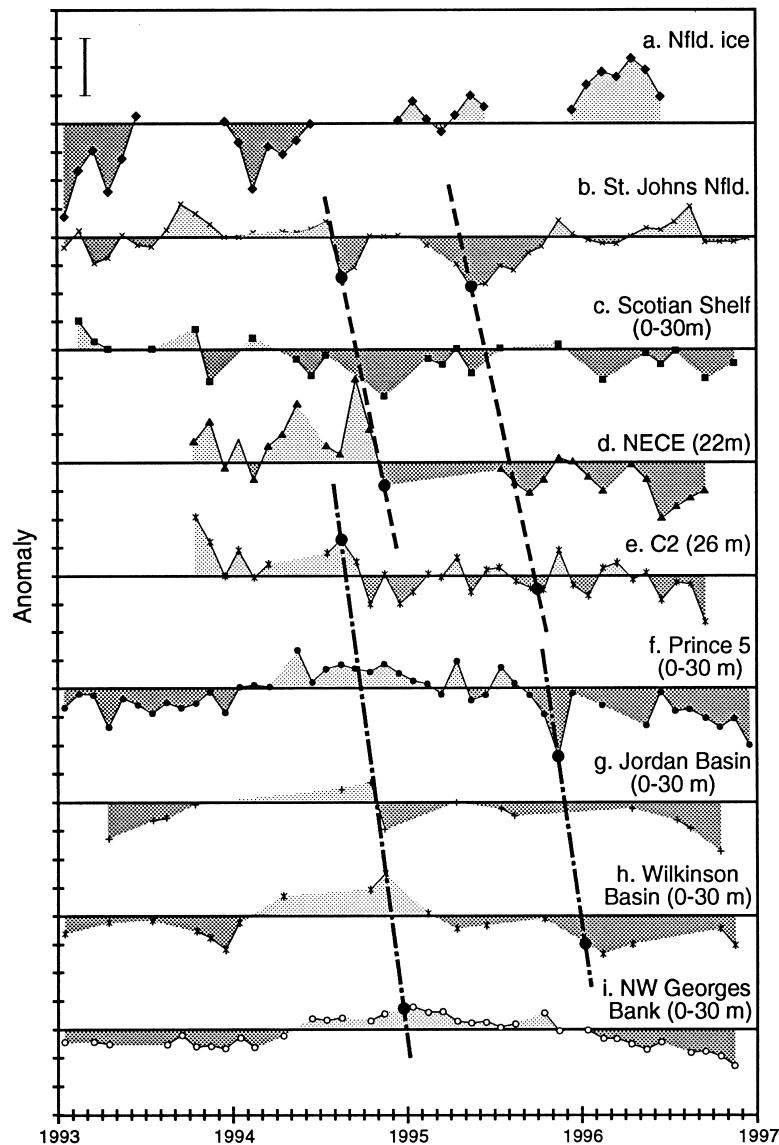


Fig. 18. Time series of monthly, near-surface salinity and (negative) sea-ice anomalies from the Labrador Shelf to the Gulf of Maine: (a) negative Newfoundland/Labrador ice cover, (b) Stn.27 surface salinity, (c) Scotian Shelf inshore (0–30 m), (d) NECE mooring site (22 m), (e) C2 mooring site (26 m), (f) Prince 5 site (0–30 m, see Fig. 1), (g) Jordan Basin (0–30 m; see Fig. 1), (h) Wilkinson Basin (0–30 m; see Fig. 1), (i) northwest Georges Bank (0–30 m; see Fig. 1). The scale at the top left stands for 1.0 salinity unit or 10^5 km^2 of sea-ice cover. Bold dashed (dot-dashed) lines indicate transit speeds of $12\text{--}16 \text{ cm s}^{-1}$ ($7\text{--}9 \text{ cm s}^{-1}$) over the outer shelf (within the Gulf of Maine).

In addition, the deviations from the annual cycles indicate that interannual variability is of comparable magnitude to decadal variability (Petrie and Drinkwater, 1993), particularly in the boundary inflows and on Georges Bank. During the first year (October 1993–September 1994) of moored measurements, enhanced deep inflow of warm, saline slope water through NEC produced

positive temperature and salinity anomalies in the interior basins of the Gulf and on the NW corner of Georges Bank. This phenomenon appeared to be associated with the presence of warm-core rings in the slope water off the mouth of NEC, but no direct link was established.

During the winter of 1994/1995, a sharp transition to a period of cold, fresh inflow occurred off southwest Nova Scotia and in the surface layers of NEC. This influx was particularly intense during 1996. As a result, surface salinities on Georges Bank decreased continuously through mid-1997, by more than 1.0 between the early months of 1995 and 1997. Oxygen isotope data indicate that the freshwater that invaded central Georges Bank is derived largely from Scotian Shelf Water (SSW) and is devoid of Maine Coastal Water (MCW). The interannual changes of these properties suggest strong variations in mixing ratio and/or volumetric inflow of the SSW, but no significant fluctuations of the SSW or slope water end points. The dynamical mechanism(s) for exclusion of MCW from central Georges Bank during 1996–1997 is(are) presently unclear.

Results of a simple box model forced by observed boundary fluxes of mass and freshwater compare well with the observed salinity changes in the interior of the Gulf and on Georges Bank. The net increase in volumetric transport through the Gulf between the first and second half of the mooring period is estimated to be roughly $10^5 \text{ m}^3 \text{ s}^{-1}$, compared to a average value of $5.63 \times 10^5 \text{ m}^3 \text{ s}^{-1}$. Comparisons with previously published estimates suggests that the SSW (deep NEC) inflow was higher (lower) during the GLOBEC field years, 1993–1996 by roughly a factor of two, but that the total volumetric transport through the Gulf is consistent with climatological mean values. The model also indicates that the average salinity of the Gulf water declines by 0.73 as a result of increased freshwater fluxes on the boundaries during the latter half of the period. This compares well with the hydrographic observations from the broadscale surveys. The net *change* in the freshwater fluxes during the latter half of the 3-yr period exceeds the climatological average value estimated by Loder et al. (1998). Finally, based on the average volumetric flow rate, the flushing time for the Gulf is estimated to be of order 1 yr.

A search for possible sources of the intense freshwater inflows during 1995–1997 indicates that local runoff and precipitation are not exceptional for this period, nor is the discharge from Gulf of St. Lawrence rivers. However, positive anomalies in Gulf of St. Lawrence and Labrador/Newfoundland Shelf ice cover and low salinities in the surface layers off St. John's, Newfoundland, suggest that the freshwater has a northern origin and is carried into the region by means of the large-scale coastal current system (Chapman and Beardsley, 1989). The key to explaining the cause of the freshening lies in an earlier event that occurred at Station 27 and in the Labrador ice cover in 1983–1985 (Fig. 17). According to Belkin et al. (1998), this event in the Labrador Sea was a manifestation of the “Great Salinity Anomaly” of the 1980s, or GSA 1980s for short. The GSA 1980s is distinguishable from an earlier advective phenomenon, known as the Great Salinity Anomaly of the 1970s (GSA 1970s; Dickson et al., 1988) in that the latter was formed in the Greenland/Iceland Seas in the mid- to late-1960s in conjunction with an excessive outflow of Arctic sea ice through Fram Strait and traced around the subpolar gyre of the North Atlantic over a period of roughly 8–10 yr. In contrast, the GSA 1980s originated in the Labrador Sea/Baffin Bay and passed around the subpolar gyre in only 6–7 yr. Its passage down the Labrador Shelf and past St. John's, Newfoundland in 1983–84 is detected in both the sea-ice cover (positive anomaly, Fig. 17f) and Stn. 27 salinity (negative anomaly, Fig. 17g). Furthermore, Belkin et al. (1998) point to evidence for another such event in the early 1990s (GSA 1990s). Like the GSA 1980s, the GSA 1990s results in excessive sea-ice cover and a twin minimum in Stn. 27 salinities in the first half of

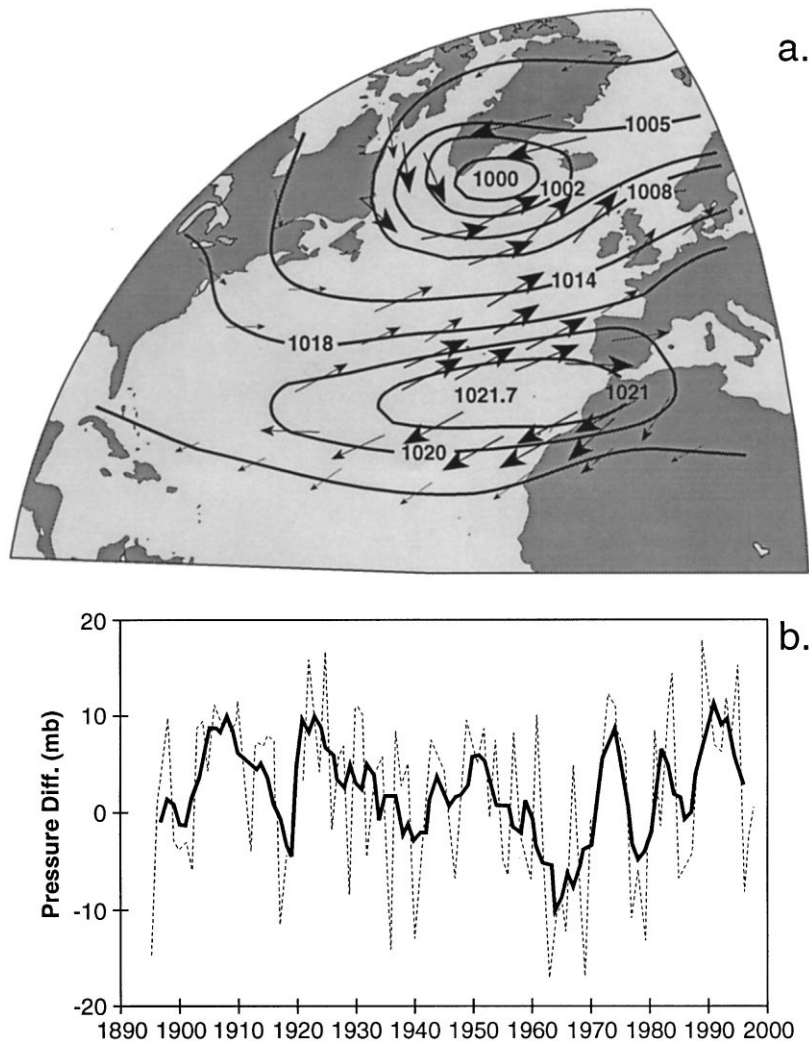


Fig. 19. (a) Climatological winter mean sea level pressure over the North Atlantic, 1961–1990. (b) Anomalies of the North Atlantic Oscillation (NAO) index as measured by the winter mean sea level pressure difference between Ponta Delgada, Azores and Iceland. Annual (dashed) and 5-yr running (solid) means are shown between 1895 and 1997.

this decade (Fig. 17f and g). Assuming reasonably constant transport to the west of Newfoundland by the inshore and offshore branches of the Labrador Current, these pulses of cold fresh near-surface waters would certainly be felt on the Scotian Shelf and Gulf of Maine after an appropriate lag for advection. However, determination of that lag is confounded by the multiplicity of pathways available and by the possibility that the transport rates are not constant. For example, Houghton and Fairbanks (2000) point to a significant annual flux of Labrador Shelf Water ($300 \times 10^3 \text{ m}^3 \text{ s}^{-1}$) through the Strait of Belle Isle and into the Gulf of St. Lawrence, where the residence time is of order 1.4 yr. Such a delay would certainly distort the picture of phase lags (3–4 months) by direct advection. Recent salinity anomaly data from western Cabot Strait show a distinct

freshening of the surface layer in 1996–1997, consistent with that observed on the inshore Scotian Shelf (Fig. 18c).

The occurrences of the GSA 1980s and GSA 1990s appear to be the direct result of local cooling and ice formation (and possibly enhanced freshwater flow through the Canadian Archipelago) caused by a series of extremely severe winters over the Labrador Sea/Baffin Bay during the early 1980s and 1990s, which are also associated with an extreme high of the North Atlantic Oscillation (NAO) index (Fig. 19b). The NAO index, measured by the sea-level pressure difference between the Azores and Iceland, represents shifts in air mass between the two major atmospheric centers of action over the North Atlantic (the Bermuda–Azores High and the Icelandic Low; Fig. 19a). During an extreme NAO high, the Icelandic Low deepens and intensifies, producing excessively strong, cold winter winds off the North American continent and over the Labrador Sea/Baffin Bay. These conditions, along with a possible contribution of freshwater through the Canadian Archipelago, lead to anomalously cold, fresh surface waters that are then transported around the subpolar gyre and west of Newfoundland to the Scotian Shelf and Gulf of Maine. The 5-yr running mean of the NAO index (Fig. 19b) shows the deep minimum in the early 1960s that is associated with the origin of the GSA 1970s, followed by a quasi-decadal fluctuation that is characterized by local maxima near the start of each of the last three decades. The most recent maxima are associated with the origin of the GSA 1980s and GSA 1990s events in the Labrador Sea/Baffin Bay; it is also reasonable to assume that severe winters occurring during the first maximum in the early 1970s reinforced the GSA 1970s as it passed through the Labrador Sea.

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