The Journal of Marine Research, one of the oldest journals in American marine science, published important peer-reviewed original research on a broad array of topics in physical, biological, and chemical oceanography vital to the academic oceanographic community in the long and rich tradition of the Sears Foundation for Marine Research at Yale University.

An archive of all issues from 1937 to 2021 (Volume 1–79) are available through EliScholar, a digital platform for scholarly publishing provided by Yale University Library at https://elischolar.library.yale.edu/.

Requests for permission to clear rights for use of this content should be directed to the authors, their estates, or other representatives. The Journal of Marine Research has no contact information beyond the affiliations listed in the published articles. We ask that you provide attribution to the Journal of Marine Research.

Yale University provides access to these materials for educational and research purposes only. Copyright or other proprietary rights to content contained in this document may be held by individuals or entities other than, or in addition to, Yale University. You are solely responsible for determining the ownership of the copyright, and for obtaining permission for your intended use. Yale University makes no warranty that your distribution, reproduction, or other use of these materials will not infringe the rights of third parties.

This work is licensed under a Creative Commons Attribution-NonCommercial-ShareAlike 4.0 International License. https://creativecommons.org/licenses/by-nc-sa/4.0/
Physical origins of Georges Bank Water

by T. S. Hopkins and N. Garfield III

ABSTRACT

The seasonal behavior of the Georges Bank Water mass is described from the context of the historic National Oceanographic Data Center file (1910-1978). The Georges Bank Water is defined as the water type most commonly found in the top 40 m of water within the 65 m isobath. Plots of its distribution show it unique within the Georges Bank/Gulf of Maine region. The uniqueness is more a result of isolation and low volume replenishment than of vertical mixing of advective inputs from adjacent stratified waters. Large scale renewal by isopycnal intrusions of Wilkinson Basin Water is only observed during late fall-early winter. These conclusions are based on an analysis of horizontal salinity gradients, on salt budget calculations, and by comparisons with T-S cycling in adjacent water masses. Horizontal salinity gradients are typically <0.2‰ per 100 km across the Bank and <0.1‰ per 100 km along the Bank, and do not reflect a mean input of fresher Scotian Shelf Water nor of saltier upper Slope Water. Evaporative heat flux is a minimum in late spring ~50 Ly/day and a maximum in early fall ~400 Ly/day. The Georges Bank Water has a salinity maximum of ~32.9‰ in early March and a minimum of ~32.1‰ in early August, making the freshening cycle shorter (5 months) than the salting cycle. The historic mean salinity is 32.56‰. Some aspects of the circulation are analyzed. An explanation for the observed jet flow along the northern flank is given from a geostrophic balance with a tidal pressure force. The jet is centered over the 50 m isobath with amplitudes ~20 cm/sec. Local convergences and divergences are created by veering of the surface Ekman transports about the perimeter of the Bank (<60 m). Steady sea level response to these transport discrepancies appears to be more effectively maintained on an integrated Bank-wide scale rather than locally. A sea-level high pressure over the Bank appears to be maintained under prevailing southwesterly winds. Its existence requires local satisfaction of continuity in the perimeter barotropic and tidally rectified flows and necessitates a lack of closure in the anticyclone circulation about the high pressure. The completeness of the anticyclonic gyre is inhibited further by the lack of deep baroclinic adjustment.

1. Introduction

Background. It is not surprising that the well defined bathymetry of Georges Bank generates distinct oceanographic conditions. The biological uniqueness of the area was not missed by the incipient American Fishery in the 17th century. Scientific
description began with geological studies in the late 1800s, and landmark biological and hydrological surveys in the first half of this century. The general behavior of the region has been described in works by Bigelow (1927), Colton (1964), and Bumpus (1973 and 1976) to which the reader is directed for historical detail.

Major findings have been the identification (Bigelow, 1927) of the seasonal anticyclonic circulation about the Bank, the northern portion of which was coincident with the southern portion of the Gulf of Maine cyclonic gyre; and the description by Clarke et al. (1943) of a unique “mixed area” over the Bank where the water mass appeared to have a relatively long renewal time. They noted that although Gulf of Maine water was the major source water for the Georges Bank mixed area water, hydrodynamic isolation preserved its identity and location even during the unstratified season. Chase (1955) reported on wind-induced water displacements, and Bumpus and Lauzier (1965) produced an atlas of monthly charts showing non-tidal currents based on drift bottle returns.

Recently, modern observational techniques and renewed research interests by U.S., Canadian, and Soviet Fisheries services, by offshore petroleum management, and by individual marine institutes have generated a wealth of new data on the Georges Bank region. The bulk of the associated analyses are still in preparation.
but have been introduced, in part, to the scientific community (e.g. WHOI, 1977; Dalhousie, 1979; Godshall et al., 1979; Magnell et al., 1979).

**Features.** Georges Bank comprises part of the outer continental shelf between New England and Nova Scotia (Fig. 1). It has been identified as a bank, at least since the time of its christening by George Waymouth in 1605. It owes its isolation from adjacent shelf areas to a sequence of geologic events culminating with the glaciation of the northeast and subsequent sea level rise. The glaciers eroding the Gulf of Maine were apparently deflected eastward around the northern edge of Georges Bank and out through Northeast Channel. During these times, the Georges Bank area received sediments either directly from the southern edge of the glaciers or from glacial outwash moving along the southern side of Georges Bank (Schlee and Pratt, 1970). The southern flank is not unlike the shelf to the southwest off New England with a shelf break at ~150 m and frequent submarine canyons (Uchupi, 1965). The northern flank has a shallower break (~40 m), a maximum slope occurring between the 50 and 200 m depths, and smoother isobathic contours. The eastern flank is a continuation of the northern flank and marks the western boundary of the Northeast Channel (230 m sill), while the western flank is only weakly
defined by a shelf trough, the Great South Channel (~70 m sill). Using the 65 m isobath to define Georges Bank, the Bank has an area of $1.66 \times 10^4$ km$^2$ and the volume of water enclosed within this isobath is 740 km$^3$, approximately 5% of the total Gulf of Maine water volume. Figure 2 gives the cumulative surface area and volume by depth. Bathymetric irregularities are more pronounced at depths less than 40 m than in the 40 to 65 m depth range. Minimum depths are 3 m at Georges Shoal and 6 m at Cultivator Shoal.

Tidal currents over Georges Bank are strongly amplified and are generally recognized as responsible for the “well mixed” nature of the water overlying Georges Bank, e.g. Bigelow (1927). Tidal current amplitudes range up to 100 cm/sec, nearly an order of magnitude greater than in the deeper off-Bank waters. The $M_2$ constituent dominates, resulting in a clockwise rotating elliptical motion: the major axis is oriented NNW-SSE with an $11 \pm 4$ km excursion and the minor axis has a $6 \pm 2$ km excursion (Haight, 1942). The spatial distribution of tidally derived kinetic energy and its dissipation have been simulated in a model by Greenberg (1977) showing a maximum in the latter of up to $10^8$ ergs/cm$^2$/cycle over central Georges and decreasing corresponding to bathymetric increases to the north and south.

The vigorous tidal circulation over Georges Bank is credited as a major factor controlling the distinctiveness of the Bank’s physical, chemical and biological properties. Over central Georges there is no summer stratified period because the tidally induced turbulence in the shallow waters causes complete vertical mixing throughout the year. Shoal depths and complete mixing minimize the vertical loss of nutrients from the euphotic zone which enables enhanced primary productivity. A result is that the Georges Bank waters retain a signature of high chlorophyll after spring and fall blooms in adjacent shelf waters have waned (Sears, 1941). Contributions to a high annual yield of primary carbon are also derived from a well recognized feature of a chlorophyll maximum coincident with the tidally mixed waters around the periphery (e.g. Sears, 1941). More accurate biological signatures are obtained from the longer lived biota; for example, Clarke et al. (1943) demonstrated that the maximum concentrations of $S. elegans$ coincided with the central Georges Bank water and were maintained in the same geographical position throughout the winter.

Since Bigelow (1927) deduced the existence of an anticyclonic flow about Georges Bank, Bumpus and Lauzier (1965) and Bumpus (1973) have substantiated and added detail to its description. Primarily on the basis of drift bottle returns, they described the “eddy” as developing during spring, surviving through summer but with some offshore veering at the eastern side, breaking down in autumn to a general southwesterly drift, and finally during winter deteriorating to a southerly drift across the Bank. Flow over central Georges is less (~4 km/day, Bumpus and Lauzier, 1965) than that about the periphery. On the northern flank Magnell et al. (1979) and Schlitz and Trites (1979) reported typical flows up to 30 km/day to the
northeast, and on the southern flank Butman (1979) found flows ~7 km/day to the southwest.

Herring spawn on Georges Bank on the gravel substratum which is maintained free of fine deposition by the high energy environment (Drapeau, 1973). Studies on the larval distributions (e.g. Bumpus, 1976) quantified the translation-dispersion of the “spawning area.” During October, a spawning patch had a birth rate that exceeded physical dispersion but was subjected to 2-15 km/day translations. The inference from these results is that the replacement time for Georges Bank waters is on the order of months.

The origin and behavior of the Georges Bank water mass has not been the subject of much comment, presumably because of its unprovocative well-mixed nature. Clarke et al. (1943) suggested it was primarily derived from Gulf of Maine waters. Hopkins and Garfield (1979) conclude the same on the basis of a two-year mean salinity of ~32.5% much closer to the surface Gulf of Maine mean of 32.4% than the adjacent offshore value of 33.7%. They also comment on the unusual consistency of the Georges Bank salinity relative to the surrounding water masses, which have been shown characteristically variable, Godshall et al. (1979). The greater shelf-slope region from the New England shelf to the Scotian Shelf has been subjected to a volumetric T-S analysis (Barinov and Bryantsev, 1972) using a USSR data file, and a water mass identification and mapping (Godshall et al., 1979) using a U.S. (NODC) data file.

2. Water mass properties

*Description of the data base.* The National Oceanic Data Center (NODC) hydrographic station data file and hydrographic data on file at Brookhaven National Laboratory (BNL) and Bigelow Laboratory for Ocean Sciences (BLOS) are used for the analysis. In November 1978, the NODC hydrographic station data file for the region 38-46N, 64-71W contained 9634 stations from 1913 through 1978. Four hundred twenty nine stations (5%) were within the Georges Bank 65 m isobath. The distribution of the remaining stations was 60% in the Gulf of Maine, 12% over the Scotian Shelf, 16% over the remainder of Georges Bank and Nantucket Shoals, and 7% in the slope water. This breakdown is approximate, done by geographic location rather than water mass distribution. Twenty-five stations were included from the BNL files and 7 from BLOS.

Table 1 gives the number of stations per month from 1913 to 1978 having temperature and salinities over Georges Bank (<65 m, see below). These are unevenly distributed seasonally with a maximum of observations in September and a minimum in January. The only major U.S. data omissions (after 1977) are those from the Bureau of Land Management, Woods Hole Oceanographic Institute, and U.S. Coast and Geodetic Survey programs, none of which were available in the NODC file by late 1978.
Table 1. Hydrographic stations within the 65 m isobath on Georges Bank.

<table>
<thead>
<tr>
<th>Year</th>
<th>Jan</th>
<th>Feb</th>
<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
<th>Aug</th>
<th>Sep</th>
<th>Oct</th>
<th>Nov</th>
<th>Dec</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>1913</td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1914</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1916</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1920</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1926</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1932</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>3</td>
<td></td>
<td></td>
<td>6</td>
</tr>
<tr>
<td>1933</td>
<td></td>
<td></td>
<td></td>
<td>3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2</td>
<td>2</td>
<td></td>
<td>9</td>
</tr>
<tr>
<td>1934</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>2</td>
<td>8</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>15</td>
</tr>
<tr>
<td>1935</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>1939</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>9</td>
</tr>
<tr>
<td>1940</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>7</td>
<td>5</td>
<td>12</td>
<td>10</td>
<td>22</td>
<td></td>
<td>56</td>
</tr>
<tr>
<td>1941</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10</td>
<td>12</td>
<td>20</td>
<td></td>
<td></td>
<td></td>
<td>42</td>
</tr>
<tr>
<td>1951</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>1954</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>5</td>
</tr>
<tr>
<td>1957</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2</td>
<td>3</td>
<td>2</td>
<td>4</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>1958</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>4</td>
<td>7</td>
<td>6</td>
<td>21</td>
</tr>
<tr>
<td>1959</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td></td>
<td></td>
<td>5</td>
</tr>
<tr>
<td>1962</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td>1964</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td>1965</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>38</td>
</tr>
<tr>
<td>1966</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4</td>
<td>4</td>
<td>4</td>
<td></td>
<td>1</td>
<td>11</td>
<td>13</td>
</tr>
<tr>
<td>1967</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10</td>
</tr>
<tr>
<td>1968</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>6</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>15</td>
</tr>
<tr>
<td>1969</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>24</td>
</tr>
<tr>
<td>1970</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4</td>
<td>5</td>
<td>21</td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td>1971</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2</td>
<td>4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>15</td>
</tr>
<tr>
<td>1973</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>6</td>
</tr>
<tr>
<td>1974</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>12</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>18</td>
</tr>
<tr>
<td>1975</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td>3</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td>19</td>
</tr>
<tr>
<td>1976</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>14</td>
</tr>
<tr>
<td>1977</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>12</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>19</td>
</tr>
<tr>
<td>1978</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>7</td>
<td>5</td>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td>25</td>
</tr>
<tr>
<td>Total</td>
<td>14</td>
<td>23</td>
<td>47</td>
<td>43</td>
<td>67</td>
<td>39</td>
<td>54</td>
<td>34</td>
<td>59</td>
<td>23</td>
<td>26</td>
<td>33</td>
<td>461</td>
</tr>
</tbody>
</table>

Definition of Georges Bank Water. The Georges Bank Water (GBW) is defined as the most frequent water type found over central Georges Bank. As a surface water mass it possesses a strong seasonal variability by virtue of air-sea transfer of heat and water vapor. In order to quantify the T-S envelope, objective criteria for its definition were developed. Discrete time periods were used to eliminate potential secular drift. The entire data file was searched for two-week periods during which four or more hydrographic stations were occupied in the area defined by the 65 m isobath. From 1934 through 1978 there were 47 instances where these criteria were met. For each time period, temperature and salinity ranges were defined by first
obtaining a depth-weighted average for the top 40 m at each station and then determining the mean plus or minus one standard deviation for each set of these depth-weighted values. Figures 3 and 4 show annual composites of temperature and salinity. The ranges were plotted at the midpoint of each two-week period and the number over each range is the year the data were obtained.

For each two-week time period, the regionally determined temperature and salinity ranges were taken to define the T-S envelope of GBW. Using these envelopes, the entire hydrographic data file for each appropriate time period was examined to determine the areal extent of GBW. Figures 5a through 5d show some of the results. The GBW has been contoured at 10 m intervals beginning with the 10 m thickness interval.

Four figures from the more comprehensive data sets are presented to illustrate

the major seasonal features of GBW distribution. The mechanisms controlling the distributions are discussed in the following sections. Figure 5a, March 1974, shows the GBW distribution at the onset of spring warming and at the early stages of the stratified season. Under the prevailing northwesterly winds, the GBW is convergent against the Slope Water. This is commonly manifested by the GBW overhanging the 65 m isobath on the southern side of Georges Bank. Another feature of this time of year is the frequent pinching off of GBW in the southwestern region.

During the stratified season, the T-S characteristics of the GBW are distinct from the surrounding water masses, Figure 5b. The GBW T-S envelope is computed...
from depth averaged temperatures and salinities, and thereby a slight stratification has the effect of limiting computed GBW volumes. The GBW distribution varies over Georges Bank, but generally is coincident with the 65 m isobath. The greatest concentration of GBW occurs along the southern flank of the Bank, either to the southwest or southeast. The lack of GBW over the shoal areas of Georges Bank is a common but not necessarily regular feature of the stratified season. It is interpreted as a flux onto the Bank of replenishment water having slightly different T-S characteristics due to a mixture of GBW and Wilkinson Basin Water (WBW).

During late fall and early winter there occurs a period when, due to more intense atmospheric buoyancy extraction and salting from convective overturning, the WBW becomes more dense than GBW. After this density reversal occurs, GBW is susceptible to massive replenishment by WBW during northwesterly wind events. When this occurs, isopycnal conditions extend from Wilkinson Basin across Georges Bank. The majority of the water with GBW T-S characteristics is found in Wilkinson Basin, and therefore is assumed to have originated there and to have extended onto Georges Bank (Fig. 5c).

Following the period when replenishment can occur, the WBW and GBW experience different T-S drifts, in part the result of late winter freshening of WBW. Figure 5d shows the GBW distribution 11 weeks later than the December 1976 distribution (Fig. 5c). There is little GBW present in the Wilkinson Basin. The portion of GBW off Cape Cod either has experienced a similar T-S drift, or it has been dislodged from the Bank as might occur during a northeasterly wind event.

**T-S cycling.** Colton (1968) provided evidence of long term temperature trends in the subsurface waters at various locations on the Gulf of Maine continental margin. These generally followed surface water trends. Hopkins and Garfield (1979) suggested a ~22-year cycle in both temperature and salinity: minima in T and S were observed in the early 1940s and the mid 1960s. The seasonal variability in temperature was too strong for the frequency of sampling among the Georges Bank data to detect any significant temperature trends. Figure 3 provides an annual composite of the depth average temperatures of GBW observations. The station-to-station variability, as indicated by the standard deviation bars, is least during the winter temperature minimum (~0.3°) and most during the summer temperature maximum (~1.5°). The “cold” years of 1940, 1941, and 1965 show lower minima than the “warm” years of 1974 and 1978.

This seasonal temperature composite was fitted with a fourth order polynomial in a least squares sense, given by the solid line in Figure 6a. A temperature minimum of 3.5°C occurs in March during the first week of spring (day 83), and a maximum of 14.6°C occurs five months later in August (day 235). Hence, the heating cycle is shorter by two months, and apparently more seasonally repetitive (Fig. 3) than the cooling cycle which is forced by a wider spectrum of events. Also
Figure 6. Smoothed fits to 40 m averaged characteristics of GBW and various adjacent water masses. a. Temperature, b. Salinity.

shown in Figure 6a are the composite temperature cycles for the surface 40 m temperature averages from the waters occupying the geographic areas boxed in Figure 1, which are intended to represent the surface water masses of Wilkinson Basin Water (WBW), Georges Basin Water (GBsW), and the upper Slope Water (uSW). WBW and GBsW are subsets of Maine Surface Water (MSW). These adjacent waters have very similar phasing. Differences in the thermal ranges are more significant. The uSW has a range of ~12.1° compared to ~9.6° for the range of WBW and of GBsW. This is interpreted as a manifestation of the vertical exchanges occurring in the southern Gulf of Maine during the stratified season (Hopkins and Garfield, 1979), and of the lateral exposure of uSW to the Gulf Stream. The GBW thermal range (~11.1°) is greater than the Gulf of Maine waters primarily because of a larger differential in the summer maxima. We expect that the GBW range more closely represents the net annual insolation, in the sense that it is isolated from advective or convective effects. However, the spring insolation on GBW may be anomalous due to local fog conditions.

The accompanying composite for the 40 m averaged GBW salinities is given in Figure 4. The station-to-station variability shows no seasonal preference and is
Figure 7. The water-type cycling for the most comprehensive time sequence of data, December 1964 to September 1966 of three water masses: The GBW, the combined WBW and GBsW, and the uSW. There were no uSW or GOM data available for the September 1965 period (point 6).

about 0.1‰ and generally does not exceed the mean annual range. The year-to-year variability is least in the spring. The most anomalous observations are those associated with the 1976-1977 winter, and perhaps the 1971 winter. The salinities were also least squares fitted along with the salinities of the adjacent water masses, Figure 6b. The GBW minimum (32.14‰) and the maximum (32.94‰) occurs after WBW by 3 and 6 weeks, respectively. The GBW freshening cycle (~5 months) corresponds with its heating cycle, preceding it by ~2 weeks. The maximum occurs at day 63 in the first week in March and the minimum at day 217 in the first week of August. The WBW maximum (33.00‰) is only slightly greater than the GBW maximum, but the minimum (31.77‰) is significantly less. The range of the GBW (0.8‰) is intermediate between the uSW (0.5‰) and the WBW (1.2‰). The GBsW curve shown is considerably damped compared to the WBW, which we have evaluated as a sampling problem, i.e. the GBsW area chosen for subsampling includes occasional intrusions of the fresher Scotian Shelf Water (SSW). The GBW and the WBW are almost completely out of phase with the uSW. These observations argue against any dominant advective control of GBW by uSW, but do suggest that a slight addition (~5%) of uSW to WBW could mix to give the GBW.

The smoothed curves of Figures 6a and 6b illustrate the mean trends of these water masses relative each to the other over many years. In this fashion the specific
T-S cycling for any given year is lost. The GBW and surface Gulf of Maine waters vary within the same historical salinity range but for any given year do not necessarily coincide in range. In Figure 7 we illustrate the water-type cycling for the most comprehensive sequence of data (Dec. 64 to Oct. 66). The GBW cycle is the most repetitive and has the smallest salinity range. Its greatest variability occurs in late winter associated with its annual salinity maximum. The Gulf of Maine water types (WBW and GBsW combined) are less repetitive and show early winter salinity maxima. The uSW is much less repetitive and is spread over a greater salinity range. The T-S drifts between GBW and uSW are different almost without exception.

Advective salt input. We attempt to quantify the advective salt input to GBW by solving a set of heat and salt balance equations. The operating hypothesis is that, if the evaporation can be properly accounted for, the amount of salt advected on or off Georges Bank can be determined, i.e.

\[ \Sigma = \Sigma_A - \Sigma_D \] (1)

where \( \Sigma \) has the dimensions %/day and the subscripts \( A \) and \( D \) represent 'advection' and 'dilution,' respectively. \( \Sigma_D \) represents a decrease in salinity when evaporation \( (E) \) is less than precipitation \( (P) \), thus

\[ \Sigma_D = \frac{(E-P)}{H} S_i \] (2)

where \( S_i \) is the initial salinity and \( H \) is the average column depth (40 m).

In order to evaluate \( \Sigma_D \), we utilize the heat balance equation for an independent expression for \( E \):

\[ Q = Q_s - Q_b - Q_h - Q_o \text{ (Langleys/day)} \] (3)

where \( Q \) is the observed heat loss, and \( Q_s \) is the incoming solar radiation. The evaporative heat loss term, \( Q_e \),

\[ Q_e = 0.007L [e_w - e_a] W = EL \] (4)

(Defant, 1961) where \( L \) is the latent heat of evaporation, \( e_w \) and \( e_a \) the water vapor pressure of water and air (mb) and \( W \) the wind factor (kts), provides the necessary expression for \( E \), but introduces the additional unknown, \( e_a \). However, \( e_a \) is also included in the back radiation term,

\[ Q_b = 1.1365 \times 10^{-7} \theta^4 [0.39-0.05 e_a] [1-0.68 C^2] \] (5)

(Berliand and Berliand, 1952 and Huang and Park, 1975) where \( \theta \) is the water temperature (°K), \( C \) is the cloud cover in tenths. The sensible heat loss term, \( Q_h \), can be computed through the Bowen’s ratio,

\[ Q_h = 3.43 \times 10^{-3} LW [\theta_w - \theta_a] \] (6)

where \( \theta_w \) and \( \theta_a \) are the water and air temperatures (°C).
The remaining variables found in these expressions were estimated from observations as follows.

\( P \) and \( C \) from Local Climatological Data (NOAA) of Block Island as representative of a maritime climate at same latitude.

\( W \) and \( Q_s \) from Local Climatological Data of Boston Airport.

\( \theta_a \) as the temperature of the Wilkinson Basin surface water on the assumption that temperature equilibration had taken place enroute to GB by the predominantly westerly winds.

\( Q \) and \( \Sigma \) from 40 m averages of GB data over the period.

A small error < 3 Ly/day is introduced by the method of solution in which a mean value of \( e_a \) in equation 5 is taken to correspond to a mean value of \( Q_s \). A larger error arises from the neglect of advective heat loss, which has a maximum of \( \sim 45 \) Ly/day in August according to the results of this section (see Figs. 6a and 8). An iterative correction for the advective heat did not seem warranted; neglecting the term is equivalent to an overestimate (<10%) of \( Q_s \). Further errors
are presumed to come from the estimates of the meteorological parameters, for example an overestimate of $Q_s$ due to local fog conditions. No data were available for assessment.

The results show a distinct pattern in the seasonal variation of evaporation. A broad minimum of $Q_e \approx 50 \text{ Ly/day}$ is reached in June. This is the period when the water vapor gradient is minimal, because the water temperatures lag the surface air temperatures, and because prevailing winds have shifted to marine southwest-erlies. As summer progresses the water warms, the water vapor gradient increases, and evaporation increases. August values of $Q_e$ were about 100 Ly/day. However, large increases in evaporation do not occur until the autumnal wind shift to north- westerlies. These continental air flows are colder and drier, and greatly increase the potential for evaporation. Their immediate cooling impact in the form of sensible heat exchange probably occurs inshore of Georges Bank within the nearshore regions of the Gulf of Maine (Hopkins, 1976). Consequently, the atmospheric buoyancy extraction processes are buffered somewhat by the heat and water vapor removed from the Gulf of Maine waters prior to arrival at Georges Bank. With strong northwesterlies, the offshore distance required for equilibration is extended to within the range of Georges Bank, Bunker (1956). For these reasons, $Q_e$ increases and becomes more variable as fall progresses. For three weeks in September 1965, $Q_e$ averaged $\sim 400 \text{ Ly/day}$, and for the entire Autumn of 1965, $Q_e$ was $\sim 300 \text{ Ly/day}$.

Vertical heat loss to depth is restricted. The composite temperature cycle (Fig. 6a) shows GBW reaching an average maximum of 2°C higher in late August than Gulf of Maine interior waters. This is attributed to the lack of vertical heat loss over the Bank. The difference persists, but diminishes to $\sim 0.5°C$ by the end of the cooling cycle. Both of the buoyancy extraction terms, $Q_e$ and $Q_s$, are enhanced slightly due to this temperature increase as the continental air flows encounter Georges Bank. This effect decreases proceeding into winter. Several estimates of $Q_e$ centered in January gave values of $\sim 200 \text{ Ly/day}$.

The results for $\Sigma_A$, the rate of advective salting, show a long spring to summer freshening cycle (March through September) and a shorter winter salting cycle. Note that this differs from the total salting (see discussion of Fig. 6b). Individual values for the periods were analyzed and a fourth order least squares polynomial fit is given in Figure 8. The $\Sigma_A$ values will reflect the error in the precipitation extrapolation, i.e. from Block Island to Georges. Block Island has fewer thunder-showers than Boston, the other data source. The summer variability in precipitation is an unavoidable error source in this analysis. The July 1970 period was an example where the short-term (14 day) precipitation averaged 3.23 mm/day. The Georges Bank salinities did not reflect this freshening, and resulted in an anomalously high $\Sigma_A = 4.6 \times 10^{-3} \text{ %/day}$. 


Table 2. Horizontal salinity gradients % per 100 km.

<table>
<thead>
<tr>
<th>Period</th>
<th>Major axis</th>
<th>Minor axis</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>Stan Dev</td>
</tr>
<tr>
<td>Dec-Feb</td>
<td>.097 ± .062</td>
<td>7</td>
</tr>
<tr>
<td>Mar-May</td>
<td>.014 ± .125</td>
<td>8</td>
</tr>
<tr>
<td>Jun-Aug</td>
<td>-.034 ± .077</td>
<td>5</td>
</tr>
<tr>
<td>Sep-Nov</td>
<td>.021 ± .064</td>
<td>6</td>
</tr>
<tr>
<td>Total</td>
<td>.028 ± .096</td>
<td>26</td>
</tr>
</tbody>
</table>

The values of $\Sigma_A$ can be used to estimate the transports from the interior of the Gulf of Maine to Georges Bank by imposing an advective balance, i.e.

$$\Sigma_A = -u\Delta S \frac{\text{area}}{\text{volume}}$$  (7)

where $u$ is the flow (positive to southeast), $\Delta S$ is salinity difference between GBW and WBW (40 m averages), the area is a 200 km x 40 m section, and the volume (600 km$^3$) is that of the top 40 m out to the 65 m isobath. Both $\Sigma_A$ and $\Delta S$ were smoothed with fourth order polynomial fits from which corresponding transports were calculated, Figure 8. The $\Sigma_A$ and $\Delta S$ are nearly out phase: When the WBW is saltier than the GBW, the GBW is gaining salt, and vice versa. The GBW salts approximately from October through February. In the Gulf of Maine this corresponds to the period of convective salting of the surface layer and high evaporation. Both of these processes are more extensive in the Wilkinson Basin region than on Georges Bank. If we assume the primary advective salt source comes from the WBW (as above), then the computed transports should be positive. During the freshening cycle, the transport averages around 4.6 km$^3$/day and during the salting cycle about 12 km$^3$/day. The zeros and discontinuities are a result of $\Sigma_A$ and $\Delta S$ passing through zero.

**Horizontal salinity gradients.** The station-to-station variability of GBW is indicated by the standard deviation bars of Figures 4 and 5. Does this variability have any spatial preference that would suggest origins for the salting and freshening of GBW? To address this question, we have computed salinity gradients along the major (060°T) and minor (150°T) axis directions by using 40-m averaged salinities (Table 2). For each cruise data set, as many independent estimates were made as possible within the 65 m isobath. From 1 to 4 values were estimated from each cruise. The N in Table 2 represents the number of cruises for each period analyzed. On an annual basis, both axes have slightly positive means <.03% per 100 km. The variability is appreciably larger than the mean, particularly for the minor axis, so that we hesitate to generalize any persistent trends. Broken into three-month intervals, a discernable pattern emerges. During the December-February period,
the GBW salinities decreases in the offshore direction (minor axis). This supports the concept of GBW salting during winter from the northwest, but does not support the concept of salting through exchange with uSW. During the June-August period, the minor axis gradient becomes positive, suggesting either that there exist mechanisms to mix the uSW into the GBW, or that the now seasonally less saline MSW is being introduced from the northwest. The former implies the presence of higher salinity values (i.e. large positive gradients) on the southern side. However, uSW intrusions are not necessarily implied. During the maximum observed gradient (0.60‰/100 km) in late June 1940, the entire southern 2/3 of the Bank was still occupied by a winter GBW water type (≤ 32.7‰) while the northern 1/3 was fresher (≥ 32.3‰). Just offshore the uSW had salinities ~34.5‰. The minor axis gradient excluding late June 1940 would have been 0.049 ± 0.140‰/100 km.

The December-February period has a significant positive gradient in the major axis (salinity increasing to the northeast). In fact from October to mid-March there were no instances of negative gradients. If nothing else, this argues against winter intrusions of Scotian Shelf Water (SSW), and suggests an entrance pattern for the more saline winter MSW moving onto the northeast portions of the Bank instead of entering from the southwest. For the rest of the year, the major axis gradient tended to be positive. Individual variations reflect cases of slightly different salinities between the northeast and southwest regions. For the June-August period the gradient averaged negative as a result of the late June 1940 case noted above. The gradient excluding this case would have been 0.24 ± 0.65‰/100 km.

The foregoing arguments provide evidence supporting the hypothesis that the GBW is controlled year around by MSW, that is, through slow replenishment during most of the year but with potentially strong flushing during the fall to winter period.

3. Circulation

The objective of this section is to present several mechanisms (surface Ekman response, tidal effects, frictional barotropic flow, gyre-like circulation, and sub-perimeter flow) that contribute to the formation and the maintenance of a unique GBW mass. The major observed circulation characteristics are comprised of an anticyclonic flow about the bank that deteriorates during winter, and a strong tidal but weak non-tidal central flow. Other treatments of aspects of GB circulation are found in Csanady (1974), Sigaev (1977), Loder (1979 and 1980), and Scarlet et al. (1979).

Surface Ekman response. As a starting point, we consult the 30 year monthly averages for wind stress and Ekman transports issued by Ingham (1977) and computed after Bakun (1973) from atmospheric pressure fields at 42N and 66W. The stresses are vector averaged from six hourly observations (Fig. 9). April and Sep-
Figure 9. The 30-year monthly vector averages for wind stress (upper) off northeastern Georges Bank. The accompanying volume transports (lower) normal to the major and minor axes of Georges Bank, deep water Ekman transports (solid) and shallow water Ekman transports (hashed).

November particularly show a diminished net stress due to the seasonal transitions between northerly and southerly components in the prevailing westerly winds. Data analysed from a further inshore point at 69W (Southwestern Gulf of Maine) revealed that the offshore location had stronger (~4%), less variable (~18%), and cyclonically veered (~5°) mean wind stress values.

These stresses can be converted to transports using the Ekman deep-water relation. We illustrate these as transport components normal to the major axis (200 km at 60°T) and to the minor axis (80 km at 150°T). They are the solid bars in the lower portion of Figure 9. The information is presented in units of km³/day for reference to the Georges Bank volume.

Over Georges Bank shoaling depths invalidate the assumptions of the deep water transport relationship. The high turbulent kinetic energy makes the transfer of momentum more efficient and has the effect of distributing the resulting horizontal movement over a greater depth. Thus over Georges Bank the probability of some horizontal momentum surviving to the bottom is increased, in which case, the transport has a downwind component. We refer to this phenomena as 'downwind veering' of the Ekman transport. In addition, the occasion of some momentum surviving to be dissipated on the bottom has the effect of reducing the net volume transport relative to the deep water situation.
To estimate the downwind veering over Georges Bank, we must make assumptions on the magnitude of the vertical eddy coefficient ($A$). We assume that the peripheral depths, at which the water columns are sustained in a vertically mixed state, are the depths to which a turbulent boundary layer can be maintained throughout a semi-diurnal tidal cycle. Then we estimate the turbulent viscosity coefficient by the following scaling (Defant, 1961 and Batchelor, 1967):

$$(\sigma - f) u = A \frac{\partial^2 u}{\partial z^2}, \quad \text{or}$$

$$A = D_m^2 (\sigma - f)$$

(8)

where $D_m$ is the depth at which column mixing is sustained, $f$ is the Coriolis parameter, and $\sigma$ is the semi-diurnal tidal frequency. Thus we estimate $A = 1050 \text{ cm}^2/\text{sec}$ at $D_m = 50 \text{ m}$, and we take $A = 150 \text{ cm}^2/\text{sec}$ in deep water. (Note: We use a left-handed bathymetric coordinate system in which the $y$-axis, or parabath, is taken tangential to the local bathymetry and the $x$-axis, or diabath, is directed toward the Bank.) A smooth function for $A$ is generated using a similar diabathic ($x$) dependence as the depth,

$$D = 120 - 110 \tanh \frac{x}{18}$$

(9)

$$A = 725 - 575 \tanh \frac{x}{18}$$

(10)

These dependencies are illustrated in Figure 10. The steady state effect of decreasing $D$ and increasing $A$ is that of producing downwind veering in the transport, and a foreshortening of the frictional response time on the order $D^2 A^{-1}$. (Hopkins,
Tidal effects. Magnell et al. (1979) pointed out some unusual characteristics of current observations recently taken on the north flank of Georges. Among these (also confirmed by Butman, 1979) is the existence of a narrow jet over the bathymetric break flowing to the northeast at speeds around 25 cm/sec. The tidal current roses of Haight (1942) and preliminary data of Schlitz and Trites (1979), Magnell et al. (1979), and Butman et al. (1978) indicate large amplification of tidal currents over the Bank. For example, Schlitz and Trites (1979) cite a four-fold amplification between the Gulf of Maine and northern Georges. The observed diabathic tidal current amplitude \((B)\) is about 12 and 65 cm/sec (Haight, 1942) in the Gulf and on the Bank, respectively; and about 38 cm/sec at the 85 m depth on the northern flank (EGG, 1979). An estimate of the diabathic dependence of \(B\) is given in Figure 11.

The first-order continuity requirement can be expressed as

\[
\frac{\partial}{\partial x} (DB) = 0
\]  

(11)
Comparison between Figures 10 and 11 shows that the product DB is not constant in the diabath, in fact B over the Bank should be about twice its observed amplitude. We expect that this reduction in the integrated tidal momentum is due to a conversion to potential energy and/or a loss to frictional dissipation. Further we expect the conversion mechanism to be the non-linear advective terms in the equations of motion.

Huthnance (1973) gave a mathematical treatment of the problem in an application to the observed tidal current asymmetries over the Norfolk Sandbanks. In his model tidal currents over steeply shoaling bathymetry generate a mean flow in the parabath. He ascribed the Coriolis force and bottom friction to be the tidal-rectification mechanisms responsible for this mean flow generation. In a recent work (brought to our attention during review), Loder (1980) has elaborated on the Huthnance model and applied it to the Georges Bank region. Loder included the refinements of mean current-tidal current interaction, spatially varying bottom friction, and rotary tidal currents. In the case of the flanks of Georges Bank, the bathymetric length scale cause the Coriolis rectification mechanism to dominate over the friction rectification mechanism. The Loder work also identifies a significant Stokes velocity, which is generated as a result of the large diabatic gradients in the flow field, and which causes the mean Lagrangian speed to be about 2/3 of the mean Eulerian speed.

For the purposes of the present work, we have posed the problem in a simpler, slightly different manner. Consider a tidal current field given by

$$u' = B(x) \cos (\sigma t + \beta(x))$$

$$v' = C(x) \sin (\sigma t + \gamma(x))$$

This represents a tidal ellipse, for which the amplitudes (B and C) and the phase angles (\beta and \gamma) are functions of x. The \sigma is the tidal frequency. Neglecting parabathic velocity gradients, the depth-averaged equations of motion can be written

$$u_t + uu_x - fv = -\phi_x - \frac{\tau^d_x}{D}$$

$$v_t + uv_x + fu = -\phi_y - \frac{\tau^d_y}{D}$$

(13)

where u and v include both the tidal and non-tidal velocity components, f the Coriolis parameter, D the depth, \nabla \phi the barotropic pressure gradient, and \tau^d the kinetic bottom stress. The continuity relation is

$$\frac{\partial}{\partial x} (Du) = -\frac{\partial}{\partial y} (Dv) - \frac{1}{g} \frac{\partial \phi}{\partial t}$$

(14)

Because of the strong diabatic gradients in bathymetry, the left-hand side domi-
nates for meso-to-low frequencies and can be taken separately equal to zero. Multiplying (13) by $D$ and (14) by $u$

$$u_t + \frac{1}{D} \frac{\partial}{\partial x} (Du^2) - f v = - \phi_x - \frac{\tau_{yx}}{D}$$

$$v_t + \frac{1}{D} \frac{\partial}{\partial x} (Duv) + f u = - \phi_y - \frac{\tau_{yx}}{D}$$

By averaging over tidal cycles, the tidal velocity components are eliminated except for the advective terms. These have a non-zero mean as follows:

$$\left< \frac{\partial}{\partial x} Du' \right> = DBB_x + 1/2 D_x B^2$$

$$\left< \frac{\partial}{\partial x} Du' v' \right> = (D_x B C + D_B C_x + D B C_x) \frac{\sin (\gamma - \beta)}{2}$$

$$+ D B C (\gamma_x - \beta_x) \frac{\cos (\gamma - \beta)}{2}$$

These two terms can be considered as equivalent to a body force, or a tidal pressure force (TPF). (Note: there also might be a contribution involving the non-tidal and tidal velocity components in these terms, as in Loder, 1980.) In the regions where the TPF has significant amplitude, it must be balanced by other terms in the momentum equations, namely the Coriolis and bottom friction terms (see next section).

From the distributions of $D$ and $B$ given in Figures 10 and 11, the TPF can easily be calculated. The parabathic flow component ($v_1$) needed in the Coriolis term for a strict balance with TPF is shown also in Figure 11. This represents the case where no other dominant forces exist in the time-averaged diabathic equation of motion. A jet-like feature is evident; the amplitude of $v_1$ peaks at ~20 cm/sec over the 50 m isobath. It changes in correlation with the tidal amplitude, as reported by Magnell et al. (1979) and with a lag appropriate to geostrophic adjustment.

We make no attempt to specifically evaluate TPF$^y$. As can be seen from (16) it depends on phase relationships between the $u'$ and $v'$ components and the diabathic phase gradients. An observational assessment is needed. We note that in the case of the curvature of the Georges Bank bathymetry, a contribution to the $y$ equation given by $-TPF^x \sin \delta$ would result, where $\delta$ is the clockwise angle between the $y$-axis and the parabath.

Frictional dynamics. We consider a steady state expression for the dynamics which includes friction and barotropic geostrophy,

$$ku - fv = -\psi_x + Au_{xz}$$

$$kv + fu = -\psi_y + Av_{yz}$$

(17)
where $\nabla \psi = \nabla \phi + \text{TPF}$, $A$ is the eddy viscosity, and $k_u$ and $k_v$ are linearizations of the nonlinear terms $uu_x$, $uv_x$ (due to tidal-nontidal flow interaction). Using the complex notation, $q = u + iv$, these equations combine to give

$$q_{xx} - \alpha^2 q = A^{-1} \nabla \psi \tag{18}$$

with $\alpha^2 = \frac{k + \text{i}f}{A}$. At the surface we require

$$-Aq_s|_{z=0} = \tau^s \tag{19}$$

where $\tau^s = \tau^{su} + \text{i} \tau^{sv}$ is the surface kinematic wind stress; and at the bottom we impose a slip condition such that at $z = d$

$$q_a = -\frac{\tau^d}{A} = -\frac{C_d}{A} |B + v_j| q_d = -\frac{r}{A} q_d \tag{20}$$

where $C_d (= 2.6 \times 10^{-3})$ is a drag coefficient and $|B + v_j|$ is a speed estimate based on the sum of the diabathic tidal component ($B$) and the velocity of the jet ($v_j$). The solution is sufficiently sensitive to the value of $r(x)$ to warrant approximating its diabathic dependence (Fig. 12). A more accurate estimate, or second iteration of $r$, might be advisable for a more rigorous quantification. The frictional parameters $k_u$ and $r$ differ in that $\frac{r}{C_d}$ is a characteristic speed [$LT^{-1}$] at the bottom, whereas $k$ is a characteristic horizontal internal shear [$T^{-1}$] within the water column. From the equations of motion, $k$ is scaled by setting $kD = \frac{1}{10}$ $\tau^{d_y}$ or by $k = r/10D$ where $\tau^d$ is approximated by $r\bar{v}$ with $\bar{v}$ representing a depth-averaged velocity. Even thus dimensioned, the $k$ term contributes in the non-integrated parabathic equation (17). The diabathic dependence of $k$ is plotted in Figure 12. The $k$ term can be interpreted as accounting for incomplete geostrophy due to internal friction.
Figure 13. The computed dependence of the transport coefficients: $E$, $F$, $G$, and $H$. The distances relate to depths as in Figure 10.

The solution to equation 16 is

$$q = \frac{\nabla \psi}{A \alpha^2} \left[ \frac{r \cosh \alpha z}{A \sinh \alpha d + r \cosh \alpha d} - 1 \right] + \frac{\tau}{A \alpha} \left[ \frac{A \alpha \cosh \alpha (d-z) + r \sinh \alpha (d-z)}{A \alpha \sinh \alpha d + r \cosh \alpha d} \right]$$

(21)

The first part on the right hand side is the combined barotropic/bottom Ekman flow and the second the surface Ekman flow. The vertical integral of $q$,

$$Q = \int_0^d q \, dz = U + iV$$

defines the total transport, the real part of which is the diabatic transport, and the imaginary part is the parabolic transport. These can be written as

$$U = \frac{E \psi_x + F \psi_y}{U_B} + \frac{G \tau^x - H \tau^y}{U_B}$$

$$V = \frac{E \psi_y - F \psi_x}{V_B} + \frac{G \tau^y + H \tau^x}{V_B}$$

(22)

where the coefficients $E$, $F$, $G$, and $H$ are plotted in Figure 13 using the parameter distributions given earlier. $E$ represents the ageostrophic effect of horizontal friction combined with ageostrophic veering in the bottom frictional layer. It has a
maximum amplification at about the 40 m isobath. $F$ represents the geostrophic effect reduced by bottom friction. Its dependence is dominated by $D/f$. $H$ represents the conventional Ekman transport coefficient, and $G$ is the deviation from it due to bottom friction. $G$ and $H$ combine to produce the downwind veering effect. For example, with a parabathic wind only, the total transport begins to veer from its deep water value ($U_s = \tau^g/f$) at \(~40\) m. By the 20 m depth, it veers by 45° ($V_s = U_s$) and is reduced to \(~70\)% of its deep water value.

The sensitivity of the coefficients in (22) relative to the solution parameters is given in Table 3.

These errors do not change significantly over the depth range of interest. Variations in $r$ mostly affect the ageostrophic barotropic component ($E$) and enhance the downwind surface component ($G$). Changes in $A$ also affect $G$ and to a lesser degree $E$. The solutions are not very sensitive to $k$.

For the case of the tidal jet with no sea level slope or wind, equations 22 become

\[
U_j = E(TPF_v) + F(TPF_v)
\]

\[
V_j = E(TPF_v) - F(TPF_v)
\]

(23)

If we assume that the tidal jet is independently non-divergent, we can impose the condition that $U_j \approx 0$, which is equivalent to (14) together with $U_j = 0$ at $x < -40$ km offshore. From (23) we can solve for TPF$_v$ and evaluate $V_j$. The tidal jet transport has a maximum over the 70 m isobath; it is more symmetric than the TPF$_v$ which has a shoaler maximum over the 50 m isobath. Integrated it approximates 20 km$^3$/day.

The TPF$_v$ thus derived has a negative maximum over the \(~20\) m isobath. The second term in the $U_j$ equation is a geostrophic onshore transport reduced by friction at the bottom. It compensates the offshore frictional transport of the first term, which has a maximum in the near bottom layer. Thus we have generally onshore flow in the upper layers and offshore flow in the lower layer. The bottom diabathic frictional flow is convergent (decreasing offshore) on the offshore side of the tidal jet and is divergent (increasing offshore) on the inshore side. A two-dimensional continuity requirement suggest downwelling into the bottom layer on the shoal side of the jet and upwelling on the deepside. The diabathic geostrophic flow has a
compensating pattern (divergent offshore, convergent onshore) about the TPF maximum.

The tidal jet plays an important role in the control of GBW in that the vertical mixing action of the jet acts as a diabathic barrier for any lateral intrusions, i.e., they lose their water type identity in the jet. It tends to entrain some offshore water from below the MSW (upwelled into the jet), to transport and to mix it while in the jet, and to cause a surface discharge onto Georges Bank.

**Gyre-like circulation.** A coherent circulation about the Georges Bank bathymetry could be an agent for reducing lateral exchange and thereby for preserving the water mass integrity of GBW. There exists significant analytical and observational evidence for a perimeter flow, however without proof of circumfluent coherency. There exists evidence that the GBW has a distinct water type, but as a waterbody it does not always conform to the bathymetry in such a way as to suggest it to be gyre-center water. We propose the hypothesis that the low frequency water mass signature of GBW is the result of a combination of low frequency dynamics over the center together with energetic higher frequency perimeter dynamics that are not necessarily coherent around the Bank but that similarly inhibit large diabathic transports.

A barotropic anticyclone requires a sea level elevation over the Bank. The primary candidate for generating a center confluence is through some spatial inequality in the wind driven transports and/or a negative wind stress curl. A distinction can be made between local and Bank-wide continuity requirements. The downwind veering produces local convergences and divergences with time scales characteristic of local winds. The question then concerns the degree to which sea level distorts in response to this local forcing mechanism. Maximum distortion would be analogous to the coastal situation, and the parabathic flow would display significant correlation with the parabathic wind component in the mesoscale frequency range. In the case of Georges Bank, the lack of a complete coastal boundary mitigates the coastal boundary constraint of vanishing diabathic transport. In the most severe situation (e.g. the north flank with a parabathic wind) the diabathic transport is reduced 65% from the deep water to the 15 m water situation. The concavity of the isobaths with respect to positive diabathic transport diminishes the constraint further. Preliminary evidence (E.G. G., 1979) indicates little correlation between local wind and sea level or parabathic flow.

We have calculated that local continuity satisfaction \((U = 0\) in eq. 22) for \(\tau^{sw} = 1\) dyne/cm² requires a sea level increase of 0.8 cm over the northern portion of the Bank and a maximum parabathic flow of 4.5 cm/sec over the 47 m isobath. A 1 cm set-up on the Bank corresponds to a confluent volume of \(\sim 0.15\) km³. However if we apply this wind to the south flank, we find that the local satisfaction of continuity is much less compatible with a Bank-wide setup. That is, the surface divergence
generated by the downwind veering now adds to the bottom divergence associated with the anticyclonic barotropic flow. This combined divergence is sufficiently large (>10 km³/day over the southflank) to preclude the maintenance of any sea-level elevation. It appears that if Bank-wide sea-level distortions exist at all, they must be relatively uncoupled from higher frequency wind forcing and be quasi-nondivergent.

The Bank-wide sea level potential can be discussed by cross-differentiating the depth-integrated versions of (17) combined with (14)

\[ \phi_t = \frac{g}{f} [D_x \phi_y - D_y \phi_x - \text{curl} \mathbf{\tau^b} + \text{curl} \mathbf{\tau^d}] \]  

(24)

where the \( k \) term has been dropped. As noted earlier, there is a cyclonic tendency in the mean wind. This results in a wind stress curl \( \sim 10^{-9} \text{ cm/sec}^2 \) and a divergence (\( \phi_t < 0 \)) over the Bank area of about 0.1 km³/day. Without spatial wind stress data, the effect of curl \( \mathbf{\tau^d} \) is difficult to evaluate.

The curl \( \mathbf{\tau^d} \) term has contributions from both the wind driven flow and from the barotropic flow. The former is generated by the downwind veering, but contributes on a Bank-wide scale only in the event of inequalities about the entire region. For example, with the mean July winds of Fig. 9, a 15% inequality between the transport in from the southwest and out to the northeast would generate \( \sim 0.1 \text{ km}^3/\text{day} \) convergence over the Bank.

The portion of the curl \( \mathbf{\tau^d} \) term associated with a barotropic flow does not tend to cancel on opposite sides of the Bank as does the wind-driven portion. Barotropic flow for an anticyclone is divergent all the way around the Bank; and therefore in the absence of a forced convergence, any circumfluous barotropic flow is dissipative (i.e., \( \phi_t < 0 \)). In the case of the tidal jet, it too contributes to a Bank-wide set-down unless it locally and independently satisfies the zero diabathic transport condition, as assumed above.

If the sea-level isopleths are not constrained to coincide with the isobaths, the first term on the right-hand side (with \( \phi_y < 0 \)) of equation 25 can provide a convergent balance to the divergent bottom flow. This has the interesting implication that the flow must shoal downstream as it moves around the Bank. However this condition is incompatible for the case of a complete gyre, since the sea level isopleths must close on each other and cannot with a decreasing radius of curvature relative to the bathymetry. The possibility of a partial gyre is not excluded by this argument, for example an anticyclonic feature that would be associated with a high-pressure ridge extending southwest from the northeastern sector across the GS Channel. Incomplete circumfluous flow seems indicated by discontinuities in water mass distributions about the perimeter (next section). Also, local parabathic variations in the sea level response to the downwind veering effect contribute to flow discontinuities about the Bank perimeter.
We might look at the baroclinic structure for evidence of a Bank-wide gyre. The bathymetry of Georges Bank precludes any diabathic baroclinic adjustment below ~30 m. Parabathic adjustments are only possible to 75 m, below which the Great South (GS) Channel interferes. As an indication of scale, we note that a similarly dimensioned (~100 km across at 36N and 4W) anticyclonic gyre in the Alboran Sea between Spain and Algeria has surface velocities from 60-90 cm/sec. At 500 m isobaric adjustment is not complete (Lanoix, 1974). Thus with little opportunity for deep baroclinic adjustment to an anticyclone, it is not surprising to find on either side of Georges Bank independent, but similarly inclined field of mass distributions. The density cross-section taken in August 1979 (Fig. 14) shows the 26.5 isopycnal ~40 m higher offshore to the south than inside the Gulf of Maine. This generates a cyclonic baroclinic pressure gradient about the Bank which acts as a partial adjustment to an anticyclonic barotropic pressure gradient ($\phi_0<0$). For example, a depth-averaged density difference of 0.25 $\sigma_t$ units over 40 m is equivalent to a 2 cm sea level rise.

Over most of the year the region is dominated by a trend for heavier water offshore. As the stratified season progresses more of a Georges Bank signature emerges. Figure 15 shows four distributions of the 40 m depth-averaged sigma-\( t \) values. In June 1940, Figure 15a, there are only tentative indications of a Georges Bank feature with the tongue-like distributions about its periphery. It is common to observe southwestward extensions of SSW between GBW and uSW. An anticyclonic-type signature of lighter surface water surrounded by more dense peripheral waters is only really approached in late summer-early fall when enough surface
baroclinicity is available to permit shallow adjustment. The September 1966 distribution is an example, Figure 15b.

With autumnal convective overturn, the surface layers in the Gulf of Maine become more dense through surface heat loss and upward salt flux. The trend for offshore density increase is reduced or even reversed. It is at this point of vertical and lateral density homogeneity that the GBW is most susceptible to replacement. Figure 15c shows more or less a latitudinal banding of the 40 m isobars. Wilkinson Basin and northern Georges are isopycnal, due in fact to coincident water types (Fig. 5c). Southern Georges is occupied by a lighter (warmer) water type, presumably that which is being replaced by WBW. As winter progresses, cooling ex-
tends to the uSW and the trend for heavier water offshore is reestablished. Figure 15d of February 1977 shows Georges Bank to have about the same lateral homogeneity with southern Gulf of Maine as was present in December 1976. However, this is only with respect to the baroclinic pressure. The water types are different (Fig. 5d) with WBW cooler and fresher than GBW and with some vertical structure.

Subsurface perimeter flow. One of the important concerns about the Georges Bank circulation is the extent to which Gulf of Maine waters are transported in a circum-Bank flow and discharged into the New England Shelf region. Of particular interest is the continuity of the Cold Pool water mass in the Georges Bank system. Beardsley et al. (1976), Hopkins and Garfield (1979), and others have speculated as to the origin of the Cold Pool Water (CPW) and its links to the Maine Intermediate Water (MIW).

The CPW is a subsurface water residing between the shelf front and shallower shelf mixed waters (between the 50 to 100 m isobath). It is a fairly stable and seasonally recurrent feature in the mid-Atlantic Bight. It is easily identifiable during the stratified season by a temperature minimum. We have analyzed six historical data sets to provide evidence of peripheral continuity between the MIW and CPW. Water types were plotted corresponding to the temperature minima found inside the shelf-slope front. These water types were connected in an anticyclonic sense, starting from Wilkinson Basin, around Georges Bank, and onto the New England Shelf. It is assumed that the resulting T-S drift from the source MIW water type is controlled by the degree of admixture with adjacent waters.

Two examples are presented for discussion, one from May 1976, and one in September 1965. Interpretation cannot be done conclusively, because the data were all derived from diabathic sampling grids from which the parabolic continuity of such a narrow feature cannot be necessarily demonstrated, and because the water types were observed synoptically and have not been adjusted for T-S drift of the source water during their lapsed advective times.

In May 1976, Figure 16a shows fairly consistent MIW water types along the north flank (N through K). Proceeding through the Northeast Channel the water types show some admixture of uSW (K through G). From G along the south flank all the way to A on the New England shelf the water type drifts back toward the MIW water type. The temperature minimum water between the 65 m isobath and this sequence on the south flank shows admixture of GBW (113, 134, 141, and 155). There is no obvious way by which water at G could cool and freshen while flowing to the southwest. We could assume that the F’ water was MIW which had made a cyclonic circuit from the Great South Channel; however, we prefer to interpret this situation to indicate that the temperature minimum water resident on the south flank is local remnant winter water.

Figure 16b shows the situation late in the stratified season, September 1965. The
diabathic sampling intervals were larger than in the previous example. A discontinuity occurs on the northflank before the NE Channel, from 30 to 19. The waters of 19, 16, and 34 indicate admixtures of Scotian Shelf temperature minimum
water (9-12) from across the NE Channel. The discontinuity implied by the freshening of the water type to the southwest (34 to 41) is difficult to interpret. The opportunity to invoke additions of MIW through the GS Channel is not possible, since this late in the season water type continuity through the GS Channel no longer exists. The water type is discontinuous across the GS Channel or through the 59 to 60 transect. The data sampling, however, does not preclude the possibility that the temperature minimum water turns anticyclonically at the GS Channel before contributing to the New England Shelf CPW. Another possibility is that the water types found at 41 and 17 indicate some three-way mixture of GBW, local remnant temperature minimum water, and advected temperature minimum water (as in 34 and 16).

Although the temperature minimum water type appears as a distinct and spatially consistent feature, the observed T-S drifts from the source MIW water type apparently is not controlled entirely by admixtures of uSW and GBW, as would be expected under an assumption of continuous anticyclonic flow. The two regions where this kind of T-S drift inconsistency prevailed were the eastern and southwestern flanks. In the former case, the T-S drift reversed towards a cooler and fresher mixture suggesting: (a) an intermittent flow of MIW source water, (b) SSW intrusions across the NE Channel, or (c) the presence of local remnant winter water. Along the southwestern flank, more water type consistency was observed early in the stratified season than later. At all times the T-S drift was toward a fresher, and usually cooler, mixture. Hopkins and Garfield (1979) have suggested the occurrence of MIW export through the GS Channel during spring, which together with local winter shelf water would contribute to a discontinuity as long as the local source persists. Also, there were occasions when the temperature minimum water in the GS Channel region showed strong admixtures of GBW and uSW.

4. Conclusions

The historical data file shows the waters over Georges Bank to be distinct in water type from the adjacent surface water masses. The GBW displays an anomalously low variability both with respect to long-term trends and with respect to seasonal cycles. It is primarily derived from MSW, more particularly from WBW, but with some addition of uSW, SSW, and Nantucket Shoals Water.

The GBW is defined as that water type most commonly found in the surface 40 m within the 65 m isobath. On this basis the GBW type was not found elsewhere in the region, except during late fall or early winter when greater volumes (Fig. 5c) were found in the Wilkinson Basin region. These are the occasion for massive isopycnal replacements of GBW (particularly the northwestern portions) by convectively overturned WBW. The northerly winds that accompany the winter cooling process facilitate renewal by forcing a divergence of GBW away from the MSW. The remainder of the year the GBW volume distribution takes various
shapes, commonly skewed to the south as the GBW is lost to the south and replenished from the northwest.

The heat content of GBW is controlled by sea surface exchange. Lateral advection is relatively small and vertical exchange is eliminated bathymetrically. The average heating cycle is five months at ~280 L/day giving a typical range of 3.5° to 14.6°C. The GBW summer maximum is greater and more persistent than that of the Gulf of Maine waters due to the lack of vertical heat loss. Evaporative heat losses of >300 L/day in the fall to ~200 L/day in the winter were estimated. Evaporation, precipitation, and lateral advection control the GBW salinity. In the mean, the first two tend to cancel making the latter more important. The primary advective source is the WBW. The GBW salinity range is slightly less (typically from 32.1 to 32.9‰), and the GBW maximum and minimum occur later than those of the WBW. Small contributions of SSW act to reduce the winter GBW maximum and those of uSW to increase the summer GBW minimum. Estimates of the advective flux from the northwest imply ~five month flushing time during the stratified period, and less than two months for winter flushings. The probability for winter event flushing is greatly increased during the complete isopycnal condition.

The distinctiveness of the GBW is definite but not so great as to suggest it owes its isolation entirely to some baroclinic mechanism. In fact the density fields (Fig. 15) generally show only some coincidence physically to the morphology of the Georges Bank region. Neither are there any strong dependencies on the atmospheric forcing from which the region might derive its uniqueness.

Although the wind field is relatively homogeneous, the resulting transport field is not due to differences in the efficiency of vertical momentum transfer on and off the Bank. The result is a sequence of transport-induced convergences and divergences about the Bank. For example, for southwesterly winds a convergence is generated against the northflank and a divergence off the southflank. There apparently are no local sea level adjustments to these convergences and divergences, the resulting barotropic flows being too divergent or unstable about the Bank. However, they may contribute to a Bank-wide sea level distortion due to slight transport imbalances integrated over the entire Bank. Only a relatively small imbalance is required, for example, a confluent volume of 0.15 km³ is equivalent to an average set-up of 1 cm over the Bank. This would generate a parabolic, geostrophic flow of about 5 cm/sec.

The stability of a sea level high over the Bank requires the condition that the sea level decrease clockwise about the Bank (ϕ<0). This precludes a symmetric, gyre-like feature with continuous flow about a central high pressure, but suggests a less stable, ridge-like feature with isobars tending to follow the bathymetry. As such the two channel regions, the GS Channel in particular, become susceptible to discontinuous, circum-Bank flow for situations when the isobars fail to coincide with these smaller scale bathymetric features.
Other dynamics occur about the perimeter which affect the GBW behavior. A tidal pressure force is generated by gradients in tidal momentum against severe bottom slopes. The result is a positive parabathic flow, estimated at \( \sim 20 \) cm/sec on the northflank. This flow is divergent without negative parabathic gradients in the TPF. The TPF and the barotropic pressure field combine linearly.

The combined pressure gradient flow has an important pattern: that of positive diabathic (convergent) flow in the upper layer and negative diabathic (divergent) flow in the lower layer. The accompanying vertical motion is downwelling on the shoal side of the flow maximum and upwelling on the deep-side of it, a shear that strongly distorts water property isopleths (Fig. 14). The strong vertical flow field embedded in this perimeter flow acts as a pre-mixer to the GBW. In this flow a slightly different and heavier water mass is generated which eventually is introduced onto the Bank in the surface portion of the column. The jet isopycnally draws water from between the MSW and MIW (Fig. 14). Because of the magnitude of the jet, considerable parabathic displacement occurs. For example, this water may enter south of Wilkinson Basin but not move onto the Bank until the Northeast sector. The northeast region appears to be the most common site of entry for replenishment water.

On the southflank the pattern in the vertical is different for the same winds because of a generally opposite downwind veering effect. The anticyclonic barotropic flow and the tidal jet flow are weaker, but both bottom divergent and surface convergent. With southwest winds, the tendency is for offshore movement at the surface, onshore in the pycnocline (\( \sim 20 \) m) and again offshore at the bottom. This action downwells and recirculates surface losses of GBW. The GBW owes some of its isolation along the southflank to the existence of intervening water masses: i.e., at the surface through extensions of SSW from the northeast and GS Channel water from the southwest, and at depth by the temperature minimum water comprised of local winter water, MIW, and Scotian Shelf temperature minimum water. The intervening water masses all tend to show admixtures of GBW. These differences relative to the northflank help explain the observed preference for MSW over uSW as source water for Georges Bank.

Acknowledgments. This research was supported by the Department of Energy under contract No. DE-AC02-76CH00016. By acceptance of this article, the publisher and/or recipient acknowledges the U.S. Government's right to retain a nonexclusive, royalty-free license in and to any copyright covering this paper. Additional support was given by NASA. The authors appreciate the encouragement given them during this analysis by J. J. Walsh and C. S. Yentsch.

REFERENCES


Received: 4 April, 1980; revised: 5 February, 1981.