The Middle Atlantic Bight Cold Pool: Evolution of the Temperature Structure During Summer 1979

ROBERT W. HOUGHTON
Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964

RONALD SCHLITZ
National Marine Fisheries Service, Woods Hole, MA 02543

ROBERT C. BEARDSLEY
Woods Hole Oceanographic Institution, Woods Hole, MA 02543

BRADFORD BUTMAN
U.S. Geological Survey, Woods Hole, MA 02543

J. LOCKWOOD CHAMBERLIN
National Marine Fisheries Service, Narragansett, RI 02882

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ABSTRACT

Temperature data spanning the entire Middle Atlantic Bight (MAB) during 1979 are used to study the structure and evolution of the cold pool. The Nantucket Shoals and New England Shelf appear to be the source of the coldest water found in the MAB in late winter. During the spring and summer, water within the cold pool in the New York Bight north of Hudson Canyon remains colder than any shelf water either to the northeast or southwest. Thus the coldest cold-pool water persists there as a remnant of winter-cooled water rather than being replenished by a colder upstream source, and south of Hudson Canyon, cold-pool temperatures decrease in June and July as colder water from upstream is advected southwestward along the coast. Both temperature data and direct current measurements suggest that the mean alongshore current has a minimum between Nantucket Shoals and Hudson Canyon. The alongshore variation of shelf topography appears to be responsible for the spatial variation in both the alongshelf drift speed and the thermal structure of the cold pool.

1. Introduction

The "cold pool" is the colloquial name given to a distinctive band of cold bottom water located over the mid- and outer shelf, extending from Georges Bank to near Cape Hatteras in the Middle Atlantic Bight (MAB) (Fig. 1). It first appears when vernal warming stratifies the surface water (Fig. 2) and persists throughout the summer until the thermocline deepens as the water column overturns and mixes in the fall.

This thermal feature was first described in detail by Bigelow (1933), who claimed that since the cold pool within the MAB was surrounded by warmer water it was a remnant of the winter-cooled shelf water and not replenished during the summer. Subsequent data (Colton *et al.*, 1968; Whitcomb, 1970; Hayes, 1975; Davis, 1979) all showed the coldest cold pool water to be located in the New York Bight (NYB), the region between Cape May and Montauk Point, but the spatial and temporal resolution was not sufficient to establish whether a connection with another cold-water source was maintained throughout the summer. More recently, Han and Niedrauer (1981) documented the advection of cold-pool water through the NYB during April–July, which they speculated originates from the Gulf of Maine in January.

The only potential sources of cold water in the summer are the Gulf of Maine and the Scotian Shelf, but access to the NYB cold pool is restricted by the Nantucket shoals and Georges Bank (Fig. 1) where vigorous tidal mixing reduces the vertical stratification in water shallower than about 40 m. There are two pathways for unmixed cold bottom water: through the Great South Channel (GSC), with a sill depth of 75 m, or from the northeast peak of Georges Bank

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and the Northeast Channel region via the southern perimeter of Georges Bank. Limeburner *et al.* (1978) show that in August 1976 the cold pool is a continuous feature near the shelf break from the northeast corner of Georges Bank into the MAB south of New England. Using historical data, Hopkins and Garfield (1981) observed a cooling and freshening of the cold pool along the southern flank of Georges Bank from east to west into the NYB which they interpreted as evidence of the presence of local remnant winter water.

Direct current measurements reported by Beardsley *et al.* (1976) and Butman *et al.* (1982) show a mean southwestward current of about 5 cm s^{-1} over the mid- and outer shelf from Georges Bank into the MAB. Because of this non-zero mean alongshelf current in the MAB, the mass of cold-pool water cannot be stagnant throughout the summer. For such a current the transit time from 69°W to the Hudson Canyon is approximately 70 days so that the cold-pool water in the NYB during July could have originated in the Gulf of Maine in June and been advected there via the GSC. An additional 70 days or so would be required for transport of Gulf of Maine water to the NYB via the Northeast Channel. Direct current measurements in the GSC (Butman *et al.*, 1982) indicate that with the possible exception of the western edge of the channel in the late winter, flow is into the Gulf of Maine. Likewise, hydrographic data (Hopkins and Garfield, 1979, 1981; Schlitz *et al.*, 1977) suggest that with the possible exception of early spring there is no contribution of undiluted Gulf of Maine water southward through the GSC. Therefore, the GSC is an unlikely pathway for cold bottom water into the NYB.

It is important to keep in mind the distinction between the source of NYB shelf water, and the source and the spatial structure of the cold pool. Because of the mean southwestward alongshelf flow, NYB shelf water originates primarily in the Gulf of Maine and Scotian Shelf with admixtures of slope water and river runoff. Cold-pool water is a water type that results from the winter cooling of mixed shelf water, and, with the formation of the summer thermocline, the cold pool becomes a distinctive bottom thermal feature that is easy to track as a continuous band from Georges Bank to near Cape Hatteras. The core or minimum temperature within the cold pool is not a conservative feature, however, nor is it uniform in the alongshelf direction as this band of relatively cool water warms and diminishes in volume as the summer progresses. The question then is to what extent the persistence of the coldest cold-pool water usually found in the NYB is the result of renewal from the Gulf of Maine or the Scotian Shelf, the only sources of colder water, and to what extent is simply due to isolation from local heating and mixing.

An unusually large number of field programs in the MAB and Georges Bank during 1979 provide the most complete data set yet available to describe the annual cycle of the cold pool. In this paper we investigate the source, evolution and alongshelf structure of the cold pool from an analysis of its temperature structure. In an accompanying paper by Ou and Houghton (1982) these results are compared with those of a simple kinematic-model calculation.

2. Data

The combined data set consists of repeated hydrographic sections and time series data from moored instruments spanning more than 800 km along the MAB shelf from March to October 1979. Our analysis is based on the following sources of data (Fig. 1).

1) In their Nantucket Shoals Flux Experiment (NSFE79), the Woods Hole Oceanographic Institution (WHOI), National Marine Fisheries Service (NMFS), U.S. Geological Survey (USGS), and the University of New Hampshire deployed an array of current meters (denoted by dots in Fig. 1) across the shelf south of Nantucket Island for a year beginning in March 1979. All current meters were VACM's equipped with temperature sensors. All available hydrographic sections made along the current meter array were compiled.

2) A Lamont-Doherty Geological Observatory (L-DGO) program to study the shelf/slope frontal regime (SWIG) conducted three 2-week cruises near the shelf break south of New England. Hydrographic sections designated as SWIG were made west of NSFE79, while those made along the moored array were incorporated into the NSFE79 data set.

3) The Northeast Fisheries Center of NMFS made six hydrographic surveys over the continental shelf from Cape Hatteras to Nova Scotia in 1979 as part of the Marine Resources Monitoring Assessment and Prediction (MARMAP) program. Stations from these surveys were grouped into six cross-shelf sections designated A–F in Fig. 1.

4) The Atlantic Environmental Group (AEG) of NMFS collected XBT data at least monthly across the shelf along the Hudson Canyon (NODC, 1980; Crist and Chamberlin, 1981) from a ships-of-opportunity program. Additional temperature data were collected intermittently along 71°W (Hughes and Cook, 1981).

5) The USGS deployed in March 1979 bottom tripods and adjacent moorings at N2 in the NSFE79 array, at Site B (73°39′W, 38°43′N) in the NYB off the New Jersey coast, and at Site K (67°34′W, 41°02′N) on Georges Bank. Water depths at these two sites was 60 m. On Georges Bank a mooring at Site A (67°24′W, 40°51′N) at 85 m depth was recovered in March 1979.

6) L-DGO conducted a hydrographic cruise (RA-

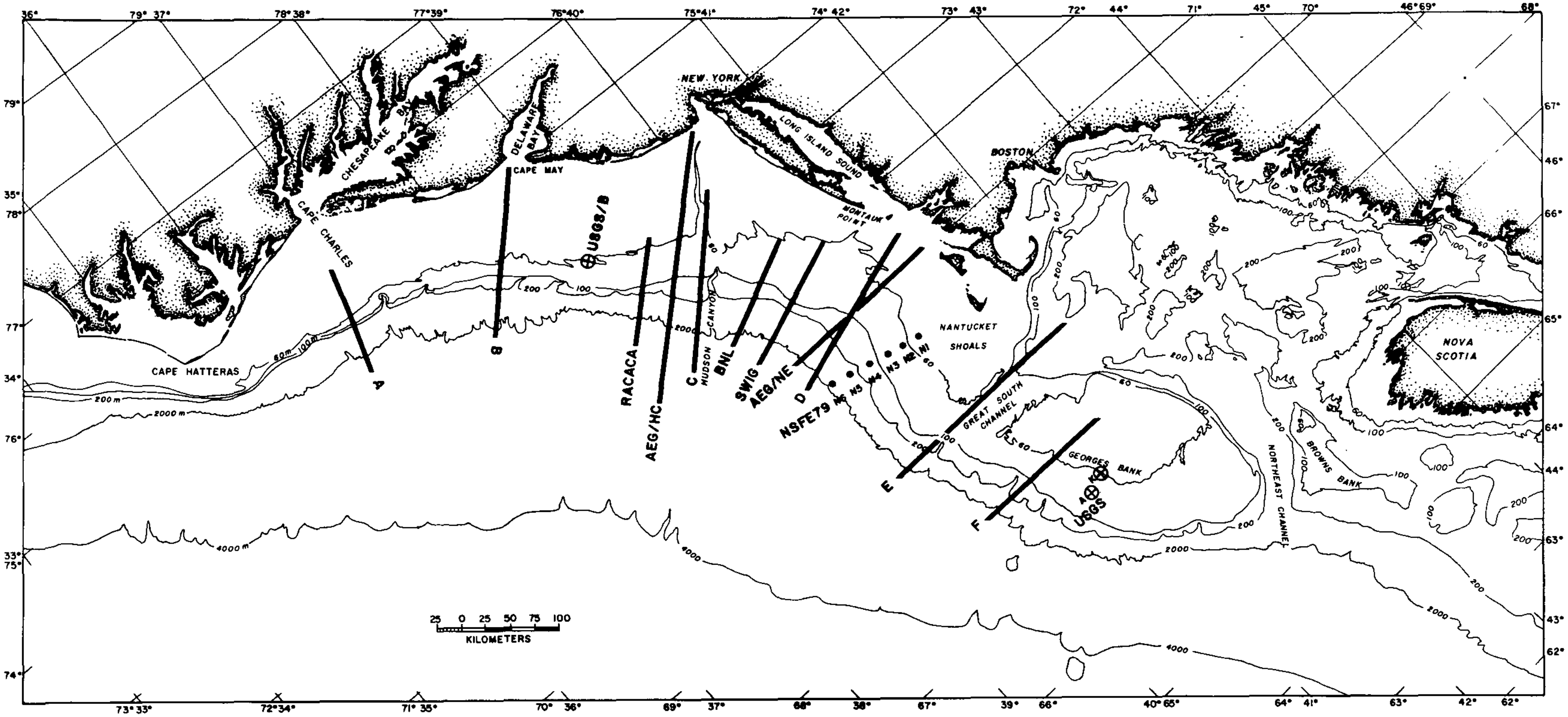

FIG. 1. Topography of the Middle Atlantic Bight, the region between Cape Cod and Cape Hatteras. Also shown are cross-shelf sections and NSFE79 moorings from which the data are derived.

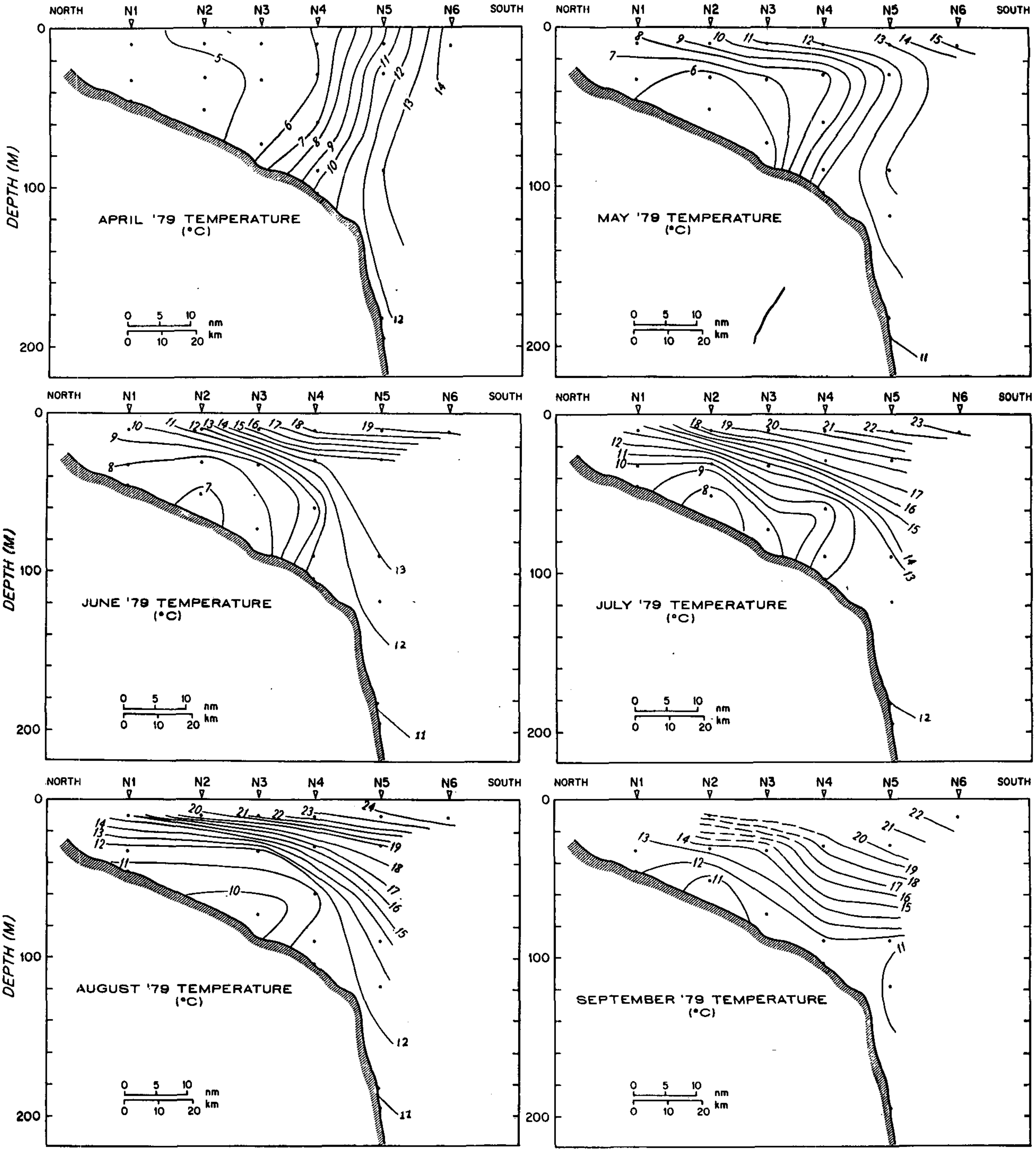

FIG. 2. Monthly averaged temperature sections along the NSFE79 array.

CACA) from south of Hudson Canyon to Cape Hatteras in June 1979.

7) The Brookhaven National Laboratory (BNL) conducted biological and hydrographic cruises in the MAB in March and June, occupying the transect south of Shinnecock, L.I., several times.

8) The NOAA Marine Ecosystem Analysis (MESA) program conducted approximately bimonthly hydrographic surveys (XWCC 21–24) from April to August containing seven equally spaced sections from Cape May to Shinnecock in the NYB.

The horizontal separation of the hydrographic and XBT stations was always less than 25 km and often less than 15 km.

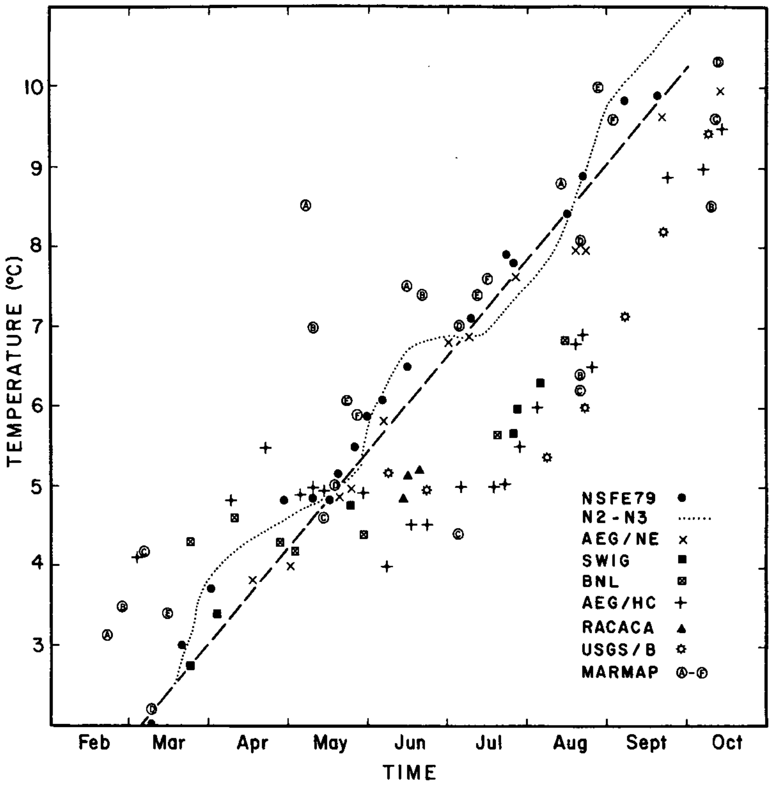

FIG. 3. Time variations of $T_{\min}$ for selected sections in the MAB. Also shown are bottom temperatures at N2 and N3 (dotted line) and USGS/B (asterisks). Dashed line approximates the warming of the cold pool on the New England shelf.

3. Analysis

We will use temperature to designate the cold pool since it is a more distinctive signal than salinity and this choice allows us to use the entire XBT, hydrographic and mooring data sets. The evolution of the cold pool is illustrated by the monthly-mean temperature sections at NSFE79 (Fig. 2). We will use the minimum temperature, $T_{\min}$, observed in a section to be representative of the cold-pool temperature. Usually this occurred near the shelf floor between the 60 and 70 m isobaths. The cold pool is a well-mixed feature 20–40 m thick surrounded by high gradients above and offshore so that $T_{\min}$ approximates the temperature of a large portion of the near-bottom water over the shelf. Because of alongshelf structure in the cold pool and discrete sampling of the data, it is possible that a single section will miss the coldest temperature in both alongshelf and cross-shelf directions so that the observed $T_{\min}$ should be considered a local upper bound. Higher temperatures do not concern us since they can always be attributed to local heating or a slope-water intrusion, whereas the presence of colder water requires explanation.

We have compiled the temperature data into the plot of $T_{\min}$ vs time shown in Fig. 3. We consider first the data from the New England Shelf (NSFE79, MARMAP D and AEG/NE). Here throughout the summer the rate of increase of $T_{\min}$ is roughly constant. The straight dashed line is hand-fitted to approximate the lowest local upper bound of $T_{\min}$. Farther southwest along the shelf in the NYB (SWIG and BNL), the warming pattern is strikingly different. After May, the rate of increase of $T_{\min}$ decreases markedly and by July $T_{\min}$ is as much as 2°C lower than (or, equivalently, lags by nearly 2 months) the $T_{\min}$ observed at NSFE79. Clearly, low temperatures in the NYB cold pool do *not* persist because of continual renewal from a *colder* source "upstream"! To the contrary, the southwest flow along the shelf must advect warmer water into the NYB, and if that advection increased the cold pool would warm more rapidly.

South of the Hudson Canyon (AEG/HC, RACACA, MARMAP C), the seasonal evolution of $T_{\min}$ is different still. During March and April $T_{\min}$ was greater than at NSFE79. $T_{\min}$ increases to a relative maximum during April and May then falls to a relative minimum in June and July before increasing again. During August and September, the lowest $T_{\min}$ was found in the southern NYB. Cooling of the cold pool here can only be accomplished by *advection*

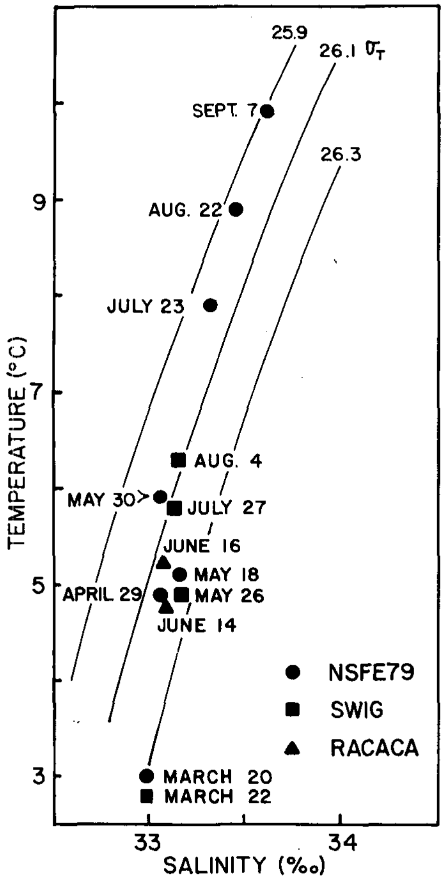

FIG. 4. T/S plot of $T_{\min}$ for selected stations. Solid lines denote isopycnal surfaces.

from a colder source "upstream" which, in this case, is the NYB north of Hudson Canyon.

The decrease of $T_{\min}$ at the Hudson Canyon (AEG/HC) from May to early June is particularly striking. The NOAA/MESA data with ~15 km station spacing provides good alongshelf resolution between the Hudson Canyon and the BNL section. During 10–15 April $T_{\min}$ was ~4.7°C throughout this region. However, during 1–6 June $T_{\min}$ decreased to ~4.1°C in a patch with alongshelf dimension of ~50 km located between the Hudson Canyon and the BNL section. The nearest source for this 4°C water is between AEG/NE and BNL at the end of April (Fig. 3). Presumably there was a patch, ~50 km long, of colder cold-pool water located just north of BNL at 1 May and just south of BNL at 1 June. The resulting southward advection speed of 2 cm s^{-1} is consistent with other estimates to be discussed later.

Two questions immediately arise: First, is the anomalously cold water in the NYB from a different source? The T/S properties of the $T_{\min}$-water (Fig. 4) show that this is not the case. In March, $T_{\min}$-water observed at both SWIG and NSFE79 have the same T/S property. As the season advances, they both evolve along the same curve which is initially isohaline gradually becoming more nearly isopycnal. After May they diverge along this curve with the water at NSFE79 warming faster.

The second question is then: Is it possible for cold water to pass undetected through the NSFE79 array? The frequency of sections along the array and the continuous low-passed temperature records from the sensors approximately 15 m from the bottom at moorings N2 and N3 located within the cold pool virtually precludes this possibility. The dotted line in Fig. 3 which denotes the lower temperature of these two records roughly coincides with the dashed line derived from hydrographic sections along the array. The individual temperature-time series from each mooring showed departures from this line consisting of pulses of water as much as 4°C warmer persisting for up to 10 days, but *never* pulses of colder water. The center of the cold pool is located between the 50 and 80 m isobaths, and since the NSFE79 array and hydrographic sections extend farther on and offshore, it seems unlikely to us that cold water can pass around the array undetected.

Although sparse, data from MARMAP sections E and F are instructive. There $T_{\min}$ always lies on or above the dashed line shown in Fig. 3. On section E, a $T_{\min}$ of 5.5°C was recorded in July, but the station was in the Gulf of Maine just north of the sill of the GSC. This water is fresher so that it would not fit on the T/S sequence in Fig. 4. Clearly, cold water from the Gulf of Maine is not passing in undiluted form through the GSC into the MAB, confirming the conclusions of Hopkins and Garfield (1981). Likewise, there is no evidence of anomalously cold water on the southern flank of Georges Bank. We conclude then that the coldest water entering the NYB is given by the dashed line in Fig. 3.

The spatial structure of the cold pool is illustrated by the 1979 MARMAP bottom-temperature maps shown in Fig. 5. Over the shelf the minimum bottom temperature and $T_{\min}$ are virtually identical. During May a relative temperature minimum in the cold pool forms south of New England. This thermal feature moves progressively southward during the summer as its temperature increases. Notice that bottom temperature increases east of the New England shelf.

In Fig. 6 isotherms of $T_{\min}$ are drawn in distance-vs-time space to illustrate the spatial and temporal structure of the cold-pool temperature changes in the MAB and along the southern flank of Georges Bank. In this figure slopes have the dimensions of velocity. We have combined NOAA/MESA and additional MARMAP bottom temperatures to obtain the largest possible data set.

The structure in Fig. 6 reflects the contribution of both local heating and alongshelf advection. In February and early March the coldest water is found on the New England shelf. Both Gulf of Maine in-

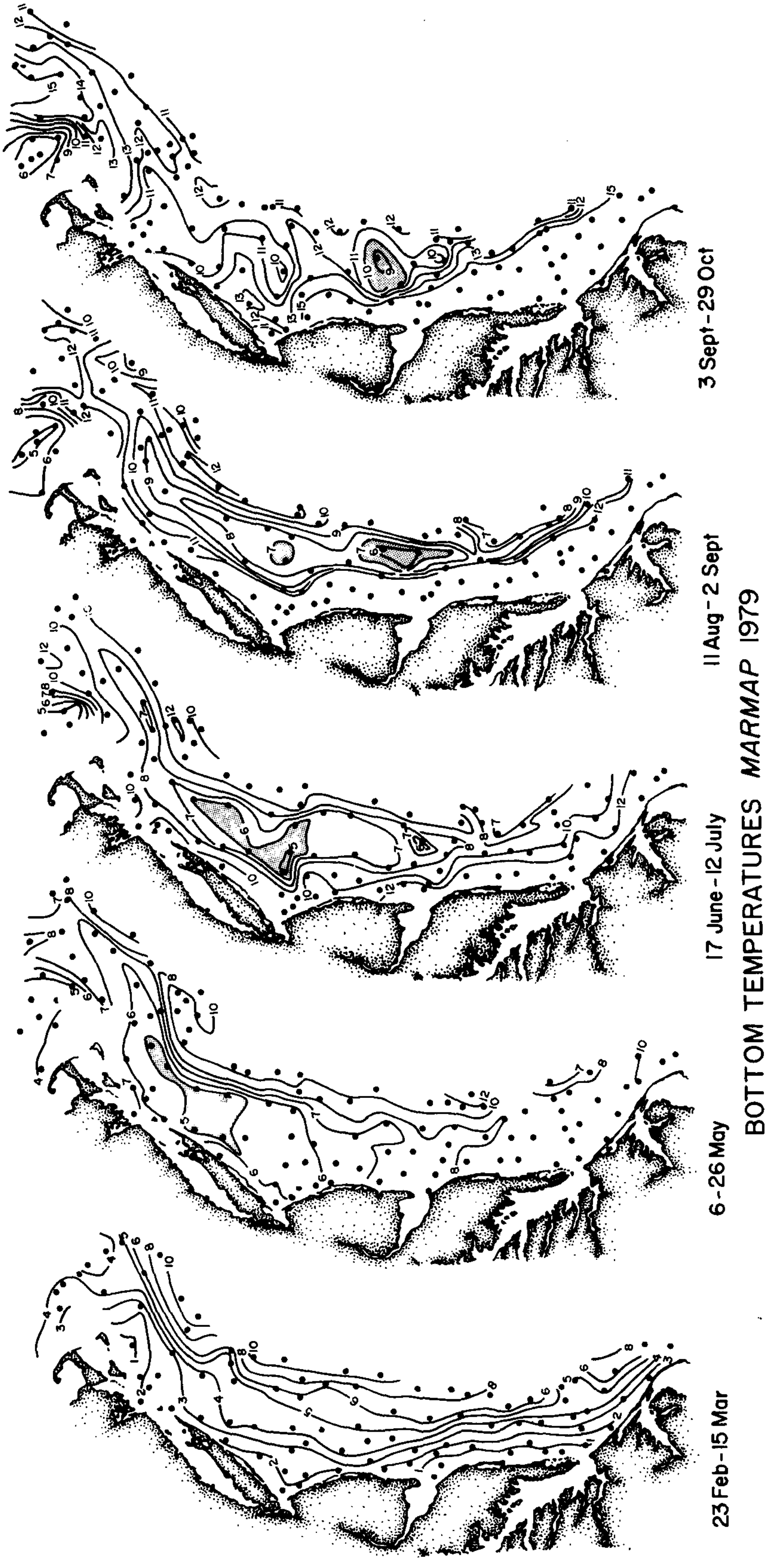

FIG. 5. MARMAP bottom temperatures out to the 200 m isobath. Contouring based on linear interpolation between stations is denoted by the dots.

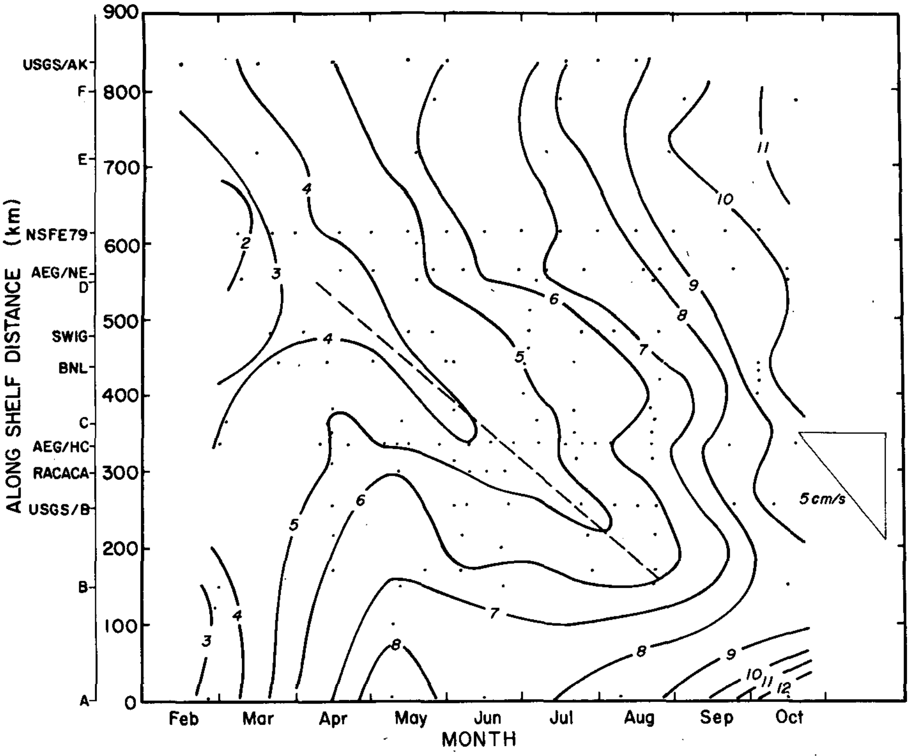

FIG. 6. Isotherms of $T_{\min}$ as a function of alongshelf distance along the 70 m isobath and time, hand-contoured using data distribution shown by the dots. The data are derived from the sections listed on left augmented by NOAA/MESA sections and MARMAP bottom temperatures (see Fig. 5). Also shown is the slope corresponding to a speed of 5 cm s^{-1}.

termediate water north of the GSC and shelf water to the east on the southern flank of Georges Bank are warmer. The sloping trough implies a southwestward drift of this thermal feature as the water warms during the summer. From the slope of the dashed line, hand-drawn through the axis of the trough, we estimate a mean drift of 3.2 cm s^{-1}.

The rate of heating of the cold pool appears to have an alongshelf variation. On the New England shelf and southern flank of Georges Bank, $T_{\min}$ increases by approximately 1°C per month throughout the summer. South of Cape May, the warming rate in March and April is ~2°C per month. The heating rate appears to be minimum in the NYB north of Hudson Canyon although the simultaneous formation and advection of the cold-pool minimum through this area makes interpretation of the data more complex. Along the axis of the $T_{\min}$ trough in the NYB, the warming is ~0.5°C per month and we take this to be a rough estimate of the warming rate of the cold pool in a Lagrangian sense.

A striking feature in both Figs. 3 and 6 is the large and persistent temperature gradient between MARMAP D and SWIG after May. Since there is warmer water upstream and no local cooling, the velocity of a water parcel cannot exceed the slope of the isotherm in this region. From the slope of the isotherms during June and July in Fig. 6, we estimate a maximum drift of 2 cm s^{-1}, which is comparable to our previous estimate. To the extent that there is local warming (due to vertical or lateral mixing), this upper bound is reduced. Since the cold pool is not confined to the bottom boundary layer, but extends at least 20–30 m above the shelf floor, the mean alongshelf drift of the water column in this portion of the shelf must have a comparable velocity. Beardsley *et al.* (1976) reported a mean alongshelf speed of 2.7 cm s^{-1} during March 1974 through a section very close to SWIG. At NSFE79, ~110 km east of SWIG, the averaged alongshelf current recorded at N2 and N3 15 m above the shelf floor during June and July was ~5.0 cm s^{-1}. At USGS/B the mean flow 15 m above the shelf floor during June was ~5.6 cm s^{-1}. Therefore, we conclude that the NYB north of Hudson Canyon is a region with relatively slower alongshelf flow.

4. Conclusions and discussion

Our primary conclusion is that the coldest water that appears as a distinct thermal feature within the cold pool in the NYB north of Hudson Canyon persists by virtue of its relative isolation rather than by

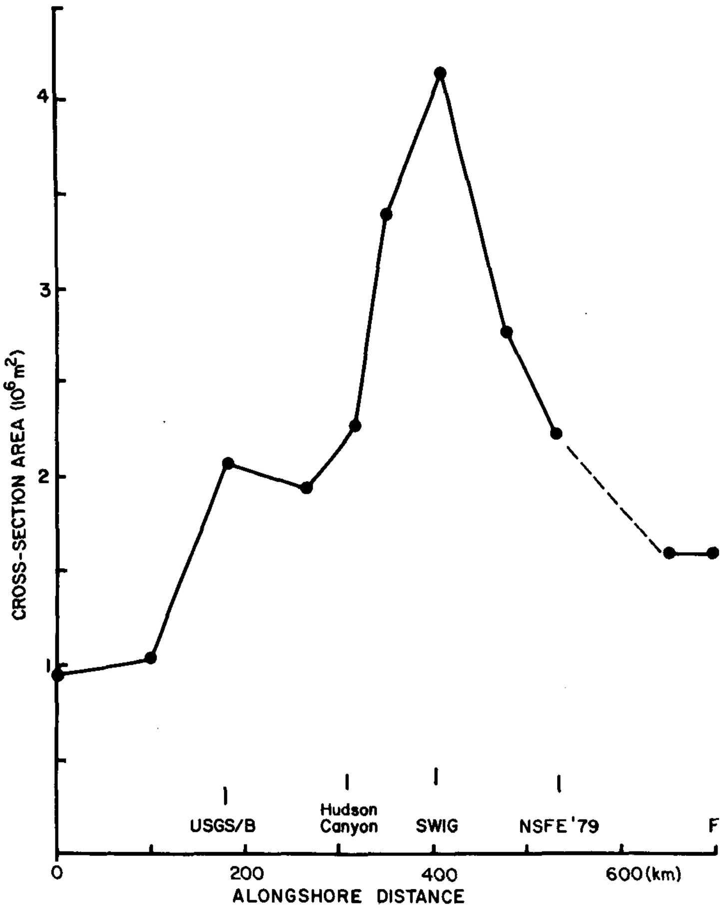

FIG. 7. Cross-sectional area A below 30 m between the 30 and 100 m isobaths in the MAB. Dashed line crosses Great South Channel where A is not defined.

renewal from a colder source upstream. Since there is a persistent and sharp temperature gradient between SWIG and NSFE79, an increased alongshelf advection to the southwest would result in warming the cold pool more rapidly here. South of Hudson Canyon the cold pool cools in June and July after rapid heating in March and April as colder water is advected into the area.

However, many interesting questions still remain. For instance, why is the coldest cold-pool water initially centered in the northern NYB near the New England shelf? The answer to this is undoubtedly related to the details of processes involved in the winter cooling and the initial heating and stratification of shelf water.

Shelf water becomes cold and well mixed in the winter on account of surface cooling and the subsequent overturning of the water column. There is little alongshelf gradient in the bottom temperatures from Nantucket Shoals to Cape Charles during early March (Fig. 5). However, in the southern MAB the cold 1°C water is confined to the very shallow shelf near the coast. Over the Nantucket Shoals high tidal energies will result in vigorous mixing down to the bottom or at least 40 m depth. This region should cool more rapidly in the winter and the subsequently denser bottom water will spread over the New England shelf and into the northern NYB creating a large volume of cold water as already noted by Hopkins and Garfield (1979).

In the same way, during the spring heat is rapidly mixed to greater depths over Nantucket Shoals. However, this water is now less dense than the surrounding shelf bottom water and will remain in the near-surface layer as it flows into the NYB. Thus we speculate that the Nantucket Shoals is the primary source of cold water for the NYB cold pool and that this source is rapidly turned off in early spring.

The evolution of the thermal structure over the shelf during the spring and summer is illustrated in Fig. 2. During early spring the initial heat input into the shelf water is mixed throughout the water column.

TABLE 1. Mean alongshore current comparing direct measurements at NSFE79 and USGS/B 15 m from the shelf bottom and inferred upper bound at SWIG with $\bar{u} = (1/A) \times$ flux assuming the flux through NSFE79 is constant along the shelf.

Location	Period	$\bar{u}$ measured (cm s^{-1})	Cross-sectional area (10^6 m^2)	$\bar{u}$ calculated (cm s^{-1})
NSFE79	June–July	5.0	2.3	5.0
SWIG	June–July	≤ 2	4.2	2.7
USGS/B	June	5.6	2.1	5.5

Then in late March or early April, the near-surface water becomes stratified. Data from the March SWIG cruise shows that this initial stratification is thermal and does not appear to be triggered by fresh water from increased river outflow. The exact timing of this initial stratification over the mid- and outer shelf is probably determined by the balance of increasing insolation and diminishing vertical mixing as winter coastal storms abate. This stratification will in turn inhibit subsequent vertical mixing of heat, and with this positive feedback the cold-pool structure will form rapidly as it becomes increasingly insulated from the heat stored in the surface mixed layer.

Throughout the summer the core of the cold pool was located between the 50 and 80 m isobaths (Figs. 2 and 5). Clearly then, this is the region of the shelf where the heating rate is minimal. There is no reason to expect a pronounced alongshelf or cross-shelf variation in the vertical flux of heat at the surface. Except near Nantucket Shoals there is no appreciable alongshelf variation in tidal energy throughout the MAB. Offshore at the shelf break, slope water is a source of heat at all depths while farther onshore enhanced tidal and wind stirring will result in greater downward mixing of surface heating. For these reasons, the core of the cold pool forms over the mid- and outer shelf.

The alongshelf variation in the structure and warming rate of the cold pool is most likely due to the alongshelf variation in the volume of water per unit length beneath the thermocline over the mid- and outer shelf. This is illustrated in Fig. 7 where we plot alongshelf variation of the cross-sectional area A below 30 m out to the 100 m isobath. A is maximum near the SWIG region and decreases rapidly south of Hudson Canyon. Therefore, at SWIG there is a greater volume of water farther removed horizontally and vertically from the adjacent heat sources. Also the same flux of heat would be distributed over a greater volume of water, in which case the cold-pool heating rate would be roughly inversely proportional to A. The consequences of this are developed in a companion paper by Ou and Houghton (1982).

The same alongshelf variation in shelf topography undoubtedly has an important effect on the magnitude of the alongshelf current as has already been suggested by Beardsley *et al.* (1976). In Table 1 we present measured and calculated alongshelf velocities assuming no cross-isobath flow. While these values are admittedly crude, the agreement is remarkable, and supports the idea that to first order the flux along the shelf in the MAB appears to be constant and the alongshelf speed is inversely proportional to A. Ou and Houghton (1982) show that this velocity variation will intensify the alongshelf temperature gradients. Thus the NYB north of Hudson Canyon is a region where a large volume of winter-cooled water resides for a longer time, resulting in the greater persistence of cold-pool water with minimum temperature there.

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