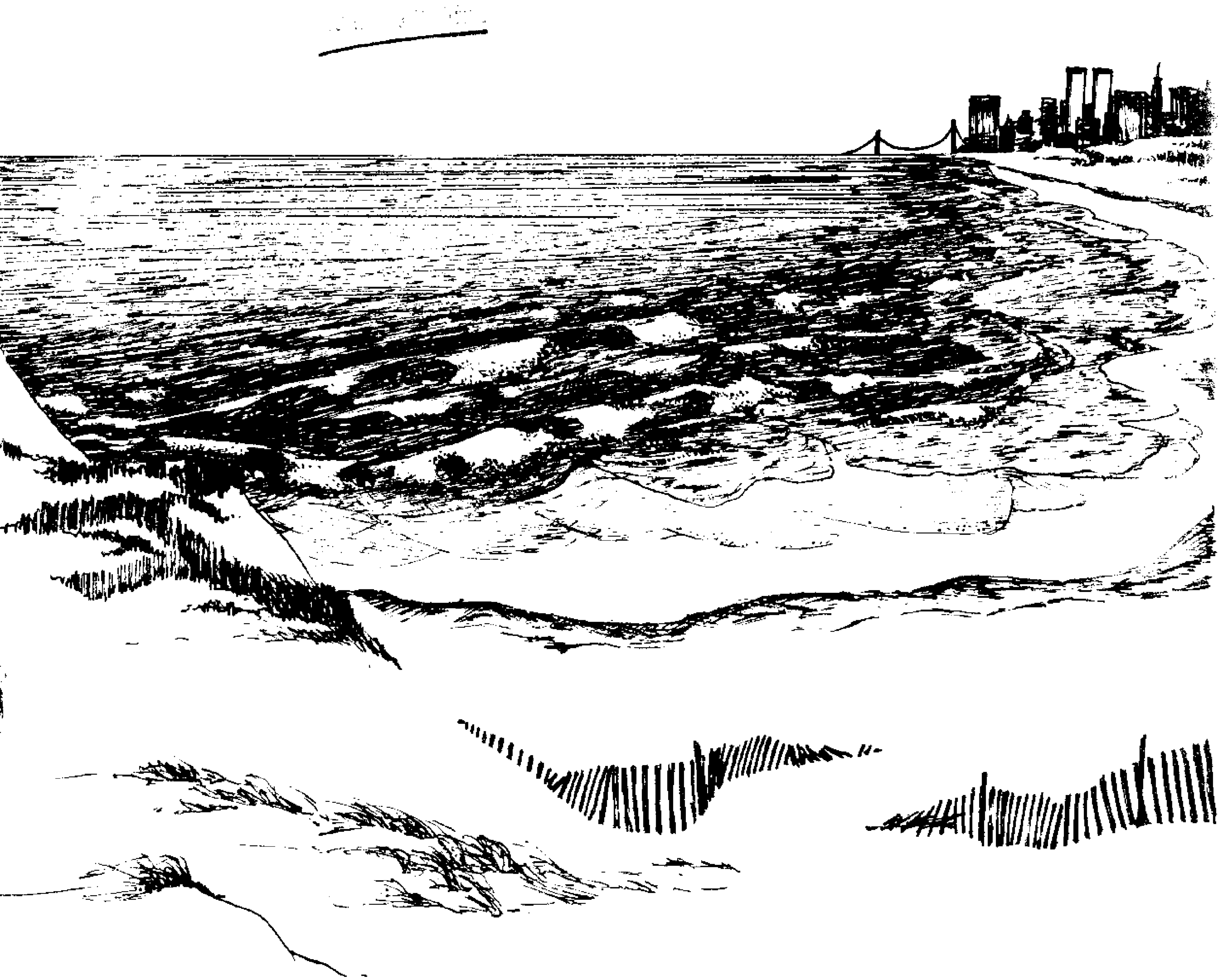
Hydrographic Properties

Malcolm J. Bowman
with cartographic assistance
by Lewis D. Wunderlich



MESA NEW YORK BIGHT ATLAS MONOGRAPH 1

The offshore water in the bend of the Atlantic coastline from Long Island on one side to New Jersey on the other is known as New York Bight. This 15,000 square miles of the Atlantic coastal ocean reaches seaward to the edge of the continental shelf, 80 to 120 miles offshore. It's the front doorstep of New York City, one of the world's most intensively used coastal areas—for recreation, shipping, fishing and shellfishing, and for dumping sewage sludge, construction rubble, and industrial wastes. Its potential is being closely eyed for resources like sand and gravel—and oil and gas.

This is one of a series of technical monographs on the Bight, summarizing what is known and identifying what is unknown. Those making critical management decisions affecting the Bight region are acutely aware that they need more data than are now available on the complex interplay among processes in the Bight, and about the human impact on those processes. The monographs provide a jumping-off place for further research.

The series is a cooperative effort between the National Oceanic and Atmospheric Administration (NOAA) and the New York Sea Grant Institute. NOAA's Marine EcoSystems Analysis (MESA) program is responsible for identifying and measuring the impact of man on the marine environment and its resources. The Sea Grant Institute (of State University of New York and Cornell University, and an affiliate of NOAA's Sea Grant program) conducts a variety of research and educational activities on the sea and Great Lakes. Together, Sea Grant and MESA are preparing an atlas of New York Bight that will supply urgently needed environmental information to policy-makers, industries, educational institutions, and to interested people.

ATLAS MONOGRAPH 1 describes the basic physical properties of the waters of New York Bight. The seasonal cycles of temperature, salinity, and density are described using three-dimensional block diagrams and planimetric maps. Bowman and Wunderlich show that the region is characterized by high variability. The seasonal temperature cycle is predictable, but the salinity shows significant year-to-year fluctuations caused by corresponding changes in river runoff and exchange with continental slope water. The hydrographic features and circulation of the outer Bight are also shown to be indirectly but significantly affected by the presence of the Gulf Stream.

Credits

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Marine EcoSystems Analysis (MESA) Program
MESA New York Bight Project

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New York Sea Grant Institute
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Lewis D. Wunderlich is completing work for his MS degree in marine environmental studies at the Marine Sciences Research Center, SUNY at Stony Brook. His research has focused on oceanography of the New York Bight apex.

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Acknowledgments

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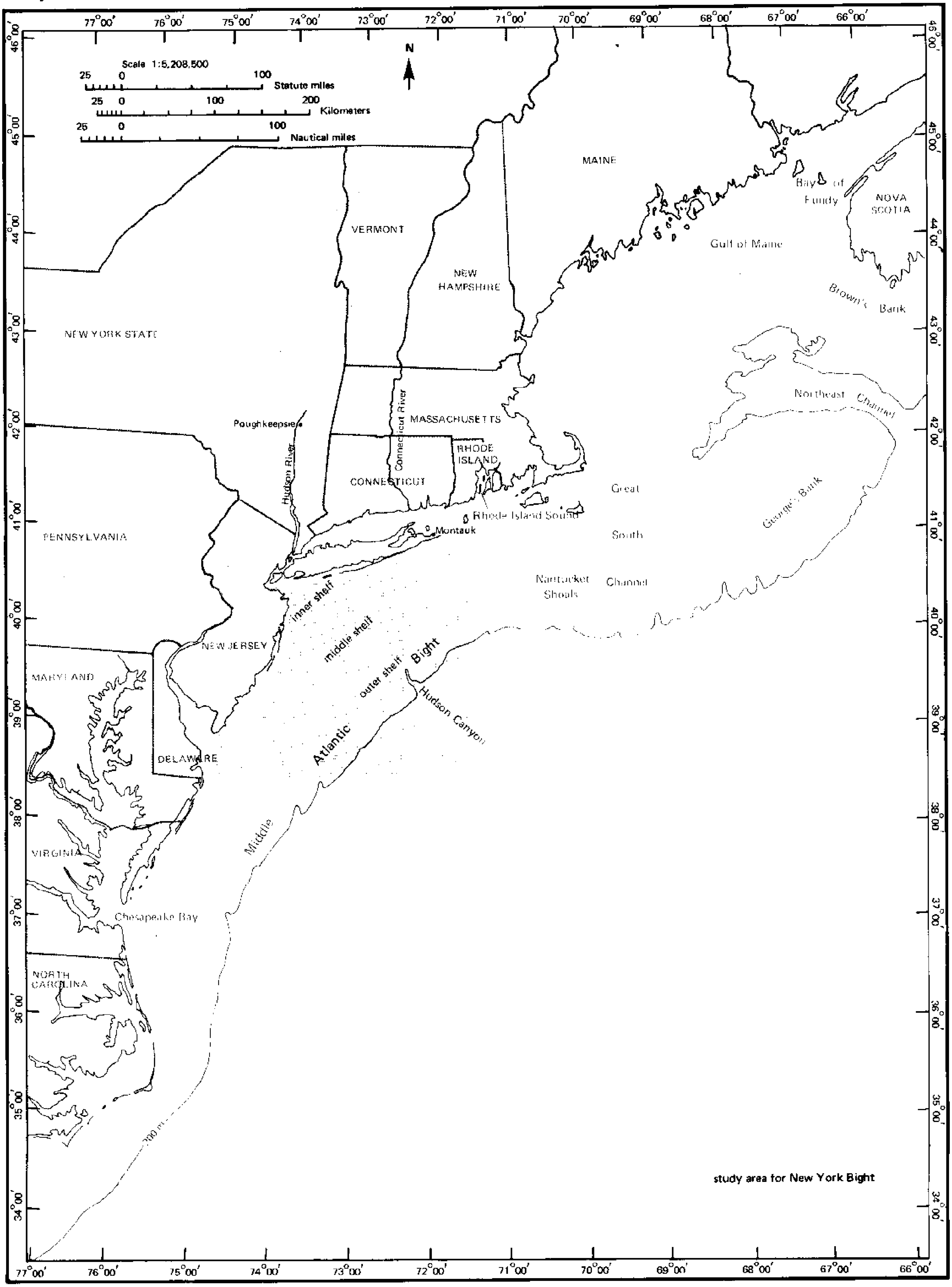
Map 1. General locator



Lambert Conformal Conic Projection

Abstract

Cycles of temperature, salinity, and density characteristic of the Middle Atlantic Bight, which includes New York Bight, are determined by seasonal patterns of insolation, river runoff, evaporation minus precipitation, winds, ocean currents, and shelf/slope exchange processes. Shelf temperatures have a large annual nearshore range around 25°C; the water column is unstratified in winter but is dominated by a strong thermocline in summer. A cool pool of bottom shelf water on the middle and outer continental shelf remains during the summer as a remnant of the previous winter and persists throughout the entire Middle Atlantic Bight.

The salinity of shelf water varies from year to year; these variations are often larger than the seasonal cycle. The principal salinity gradients are vertical and cross-shelf; salinities are lowest near shore. Salinities increase with depth on the shelf, but the vertical gradient reverses over the continental slope. Salinities are highest in winter and lowest in late summer. Total freshwater volume on the shelf equals an average of 18 months' discharge of river water.

The density field in winter is principally determined at all depths by salinity; patterns of surface density throughout the year are similar to those of surface salinity. The summer thermocline converts the winter's unstratified density field into one with strong vertical stratification.

The principal winter hydrographic feature of the outer shelf is a strong temperature/salinity front separating shelf water from slope water. During summer the thermocline eliminates any surface manifestation of the temperature front, but the salinity front persists throughout the year. A temperature/salinity inversion underlying the fronts is formed of shallow slope water and inclines shoreward, intersecting the shelf break at about 150 m.

Shelf/slope exchange processes and slope frontogenesis are dominated by small-scale dynamics, vary greatly with time and space, and are difficult to measure. Gulf Stream meanders and eddies occasionally pass through the outer Bight, creating major disruptions in the hydrographic properties of the slope waters.

Introduction

The hydrographic properties of New York Bight are complex. In addition to a seasonal cycle of temperature, salinity, and density over the continental shelf, the characteristics of slope water and its exchange with shelf water and the proximity of the Gulf Stream lead to a number of important transient phenomena.

For this monograph New York Bight is defined by 38°30′N, 71°30′E, and the coastlines of Long Island and New Jersey (Map 1). A primary source for this report was the National Oceanographic Data Center (NODC), Washington, DC. Much of NODC's data (Figure 1, Table 1) are filed by 1° *Marsden squares* (italicized terms defined in Glossary at back); these boundaries maximized the amount of data that could be used in preparing many of the figures in this monograph.

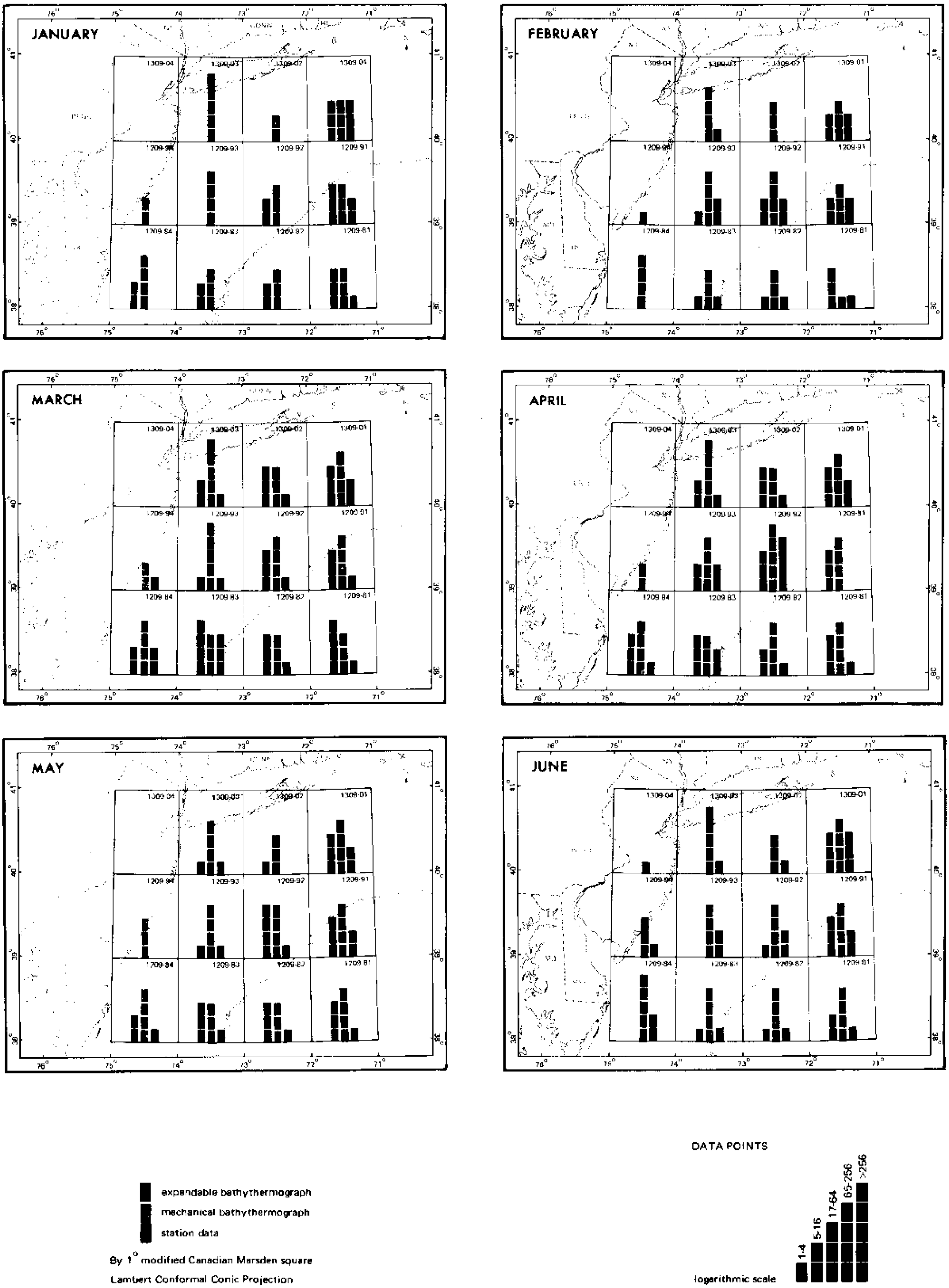
For each 1° Marsden square a weighted mean for temperature was calculated from three data sources—NODC station data, *mechanical bathythermograph*, and *expendable bathythermograph* files. Additional temperature data were taken from the *Serial Atlas of the Marine Environment* (Walford and Wicklund 1968), describing the monthly temperature structure along the US east coast. This data came from NODC files up to 1963, from US Naval Oceanographic Office standard hydrographic station data, and from the large bathythermogram trace print file at Woods Hole Oceanographic Institution.

Other useful sources were: *Bottom-Water Temperatures on the Continental Shelf, Nova Scotia to New Jersey* (Colton and Stoddard 1973); the US Naval Oceanographic Office's (1969-1974) extensive oceanic prediction studies on the northwestern Atlantic Ocean (published in *Gulf Stream*); and the classic papers by Bigelow (1933) and Bigelow and Sears (1935), presenting extensive data with interpretation of the seasonal distribution of temperature and salinity over the continental shelf from Chesapeake Bay to Cape Cod.

For this report we chose different averaging periods for temperature, salinity, and density, according to their seasonal trends and temporal variability. Within each averaging period, a simple arithmetic mean was taken for all relevant data.

In this monograph, tables at the back of the Glossary show conversions for temperature (Celsius to Fahrenheit) and distance (metric to English), though the monograph series as a rule shows both in text.

Figure 1. Data distribution from National Oceanographic Data Center files



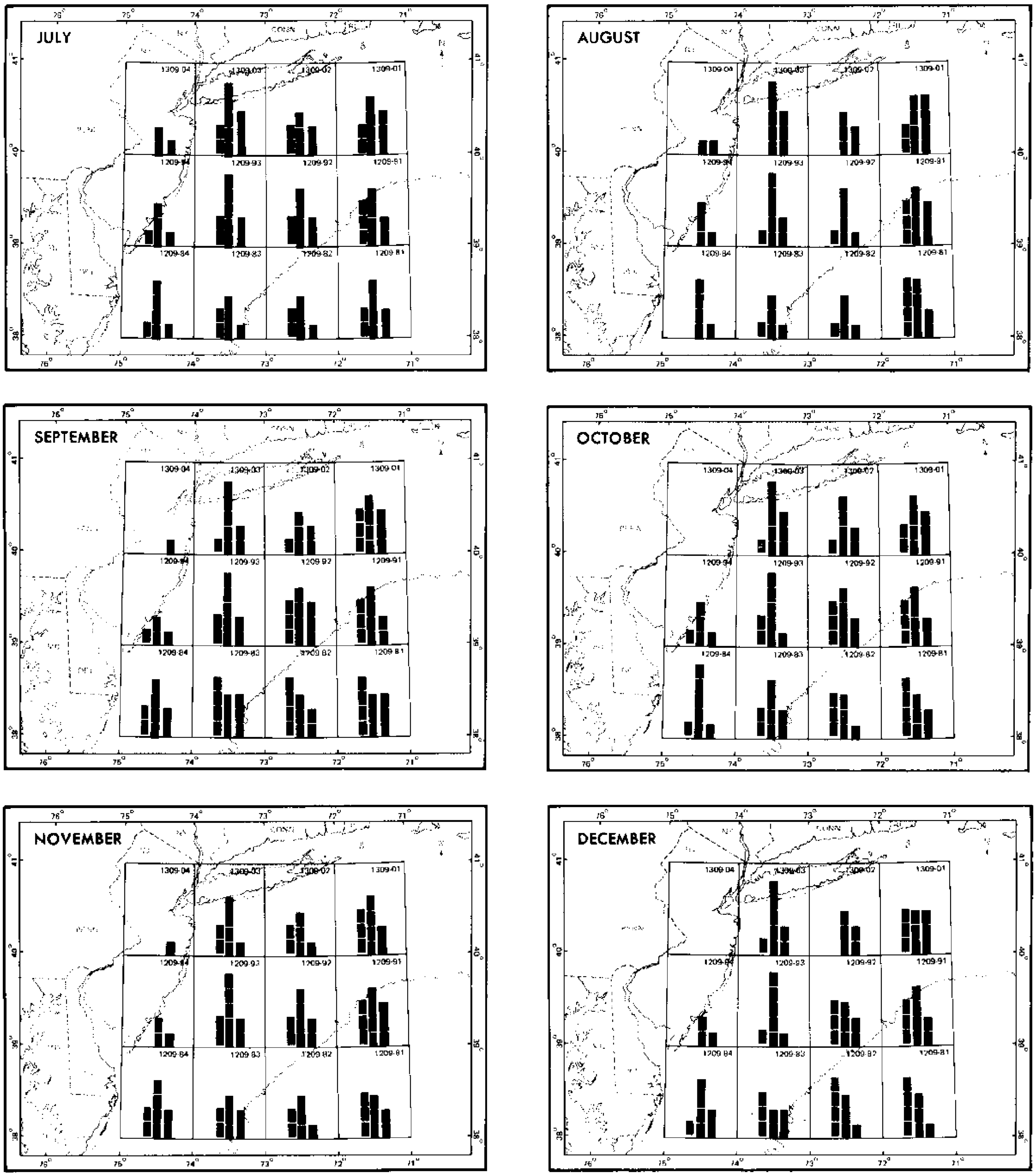


Table 1. Earliest years of NODC hydrographic data

Marsden Square	Year	Marsden Square	Year
1309-04	1949	1209-92	1920
1309-03	1931	1209-91	1916
1309-02	1932	1209-84	1927
1309-01	1916	1209-83	1931
1209-94	1931	1209-82	1927
1209-93	1928	1209-81	1927

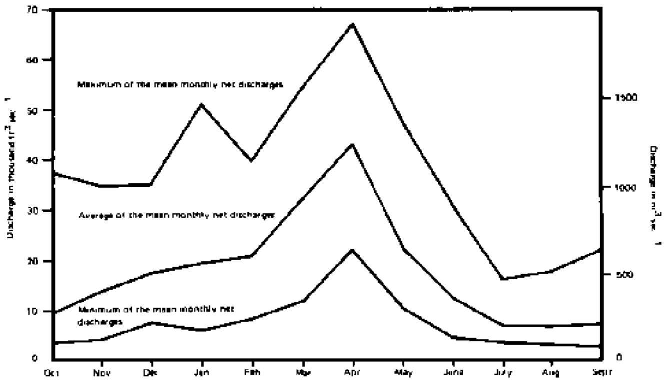
River Runoff and Local Circulation

Hudson River and Estuary

The discharges of the Hudson and Connecticut rivers, the two largest sources of fresh water in the Northeast, play a significant role in salinity distribution and inner New York Bight shelf circulation; the Delaware River (Ketchum 1952) has only minimal influence. The Hudson River drains a watershed of about 35,000 km^2 and empties directly into the Bight apex. About 50% of the annual discharge occurs between February and May (Giese and Barr 1967). The mean discharge of the Hudson River over the *water years* 1947 to 1965 was 560 m^3 sec^{-1} (20,000 cubic feet per second or cfs).

Figure 2 summarizes the mean monthly discharge at Poughkeepsie, NY. Discharge peaks in April, with a mean flow of 1,200 m^3 sec^{-1} (43,000 cfs), and is lowest in August when the mean flow is only 190 m^3 sec^{-1} (6,900 cfs). Mean monthly discharge for any given month may vary by a factor of 10 from year to year. The largest recorded mean monthly discharge was in April 1960 when the flow was 1,900 m^3 sec^{-1} (68,000 cfs) at Poughkeepsie. In 1965, due to the drought of the early 1960s, the April discharge dropped to only 700 m^3 sec^{-1} (25,000 cfs). The October 1964 mean value was a mere 92 m^3 sec^{-1} (3,200 cfs).

Hudson *estuary circulation* under mean tide and mean discharge conditions is well known from US Geological Survey work (Giese and Barr 1967; Darmer 1969; Busby and Darmer 1970), US Army Corps of Engineers hydraulic model studies (Simmons and Bobb 1963, 1965), and the work of the US Army Corps of Engineers Commission on Tidal Hydraulics (1961).



Source: From Giese and Barr 1967

Figure 2. Mean monthly net discharges of Hudson River at Poughkeepsie, NY, for water years 1947-1965

The average *tidal flux* near The Narrows is about 41,000 m^3 sec^{-1} (1.5 million cfs), about 75 times larger than the mean river discharge (Parsons 1913). The tide in the lower Hudson is dominated by the semidiurnal component and moves nearly as a progressive wave; the mean tidal range at The Battery is 1.4 m. This tidal wave enters Lower Bay and propagates upstream to the Federal Dam at Troy, NY. Superposition of the Hudson tidal wave and the Long Island Sound tidal wave leads to a *hydraulic flow* in the East River tidal strait where the currents are determined by the local *hydraulic gradient* (Bowman 1976).

For this monograph the effects of the Hudson River discharge on the distribution of hydrographic properties in the estuary were studied by summarizing properties during the two seasonal extremes in flow—the spring runoff peaking in April and the late summer minimum in July, August, or September. The Marine Sciences Research Center conducted synoptic surveys of the hydrography and circulation of New York Harbor on 8 April 1972 and 21 September 1972 (Map 2). These two days closely coincided with the extremes of the Hudson River discharge for that year: April—1,400 m^3 sec^{-1} (51,000 cfs); September—190 m^3 sec^{-1} (6,900 cfs).

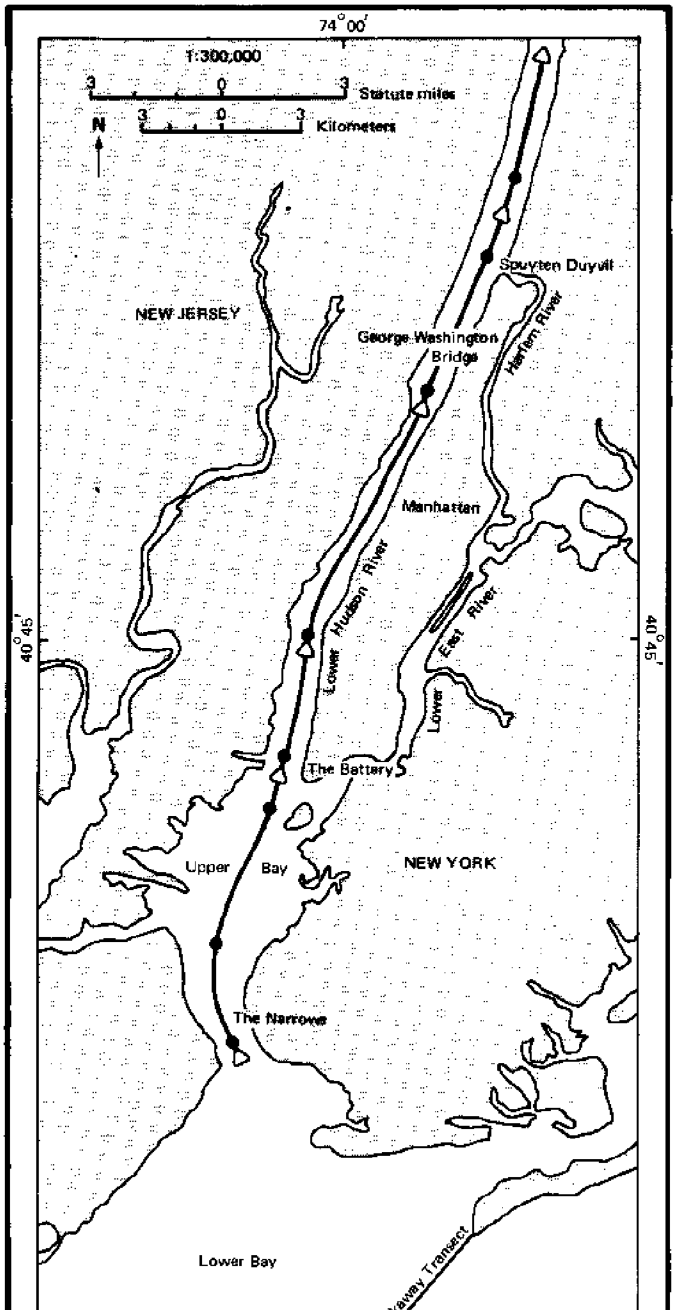
On 8 April 1972 in Figure 3, for local slack after ebb the river flow and estuarine nontidal component of circulation transport cold river water (temperature less than 4.5°C) over slightly warmer inner shelf water (temperature greater than 5.5°C). The salinity section displays the strong stratification that occurs during periods of high runoff; bottom values are greater than 25 parts per thousand (‰) at The Narrows and surface values are less than 4‰ about 40 km upstream in the Hudson River. The density distribution illustrates the intense horizontal and vertical density gradients typical of periods of high runoff.

Approximately six hours later on 8 April 1972 (Figure 3), at slack after flood, the 5.5°C *isotherm* has been moved about 10 km upstream (this approximates the *tidal excursion*); cold surface water—less than 4°C—has also been advected upstream. The salinity and density sections again illustrate high gradients and upstream tidal *advection* of the *isolines*.

Quite different temperature distributions were observed on 21 September 1972 (Figure 3), which illustrates the effect of local heating from the 185 megawatt 59th Street power plant effluent released into the lower Hudson. Surface temperatures have risen above 20°C. This warm surface core overlies Upper Bay bottom water of less than 19°C and is separated from upstream bottom river water greater than 20.5°C.

In September's salinity distribution for slack after ebb, both the slope of the *isohalines* and the salinity in the estuary have increased, compared to April. Although salinity at The Narrows is only a little more than during April observations, salinities 40 km up the river have increased to about 12°/oo.

Map 2. Hudson River locator for Figures 3 and 4



Density again is dominated by salinity; the *isopycnals* closely follow the isohalines. Density (*sigma-T* or σ_T) is greatest in Upper Bay, reaching over 17 near the bottom, and steadily decreases upstream to less than 7.

In the temperature profile for slack after flood on 21 September 1972 (Figure 3), isotherms in Upper Bay have been moved upstream and the warm core, now with surface temperatures over 21°C, has moved some 10 km upstream. The salinity regime is similar to that for slack after ebb, but the isohalines have been moved upstream through the tidal excursion. Salinity in Upper Bay is greater than 26°/oo and drops uniformly to less than 14°/oo upstream. The density profile again is determined principally by salinity; the isopycnals follow the isohalines. Sigma-T is greater than 18 in Upper Bay and drops to less than 9 upstream.

Estuarine or gravitational circulation, driven by the *horizontal density gradient,* has been investigated along a transect between Rockaway Point, Long Island, and Sandy Hook, NJ, by Stewart (1958) and in more detail by Kao (1975). Data in both cases were gathered from a 2-7 June 1952 study by the Coast and Geodetic Survey; 16 current meters moored to seven vertical strings were suspended in a linear array across the transect. Kao calculated nontidal fluxes through the section averaged over an integral number of tidal cycles during the six days. He found that a strong estuarine flow exists, with dense water flowing upstream (into the page in Figure 4) at depth under less dense near-surface water flowing seaward (out of the page in Figure 4). Upstream fluxes occur near the bottom of both the Sandy Hook and Ambrose channels; speeds are greater than 6 and 4 cm sec^{-1}, respectively. A strong upstream flow with velocities greater than 12 cm sec^{-1} appears near the Rockaway Point side of the estuary; there is no net seaward flow in this region. Surface seaward flows with speeds greater than 16 cm sec^{-1} show up near the center of the section.

The tendency of the upstream flow to be concentrated on the Rockway Point side of the section is due to channel irregularities, local accelerations associated with the curvature of the flow through Lower Bay, and intrusion of oceanic water from east of the estuary.

Kao reported that the mean nontidal fluxes through the cross section averaged over the duration

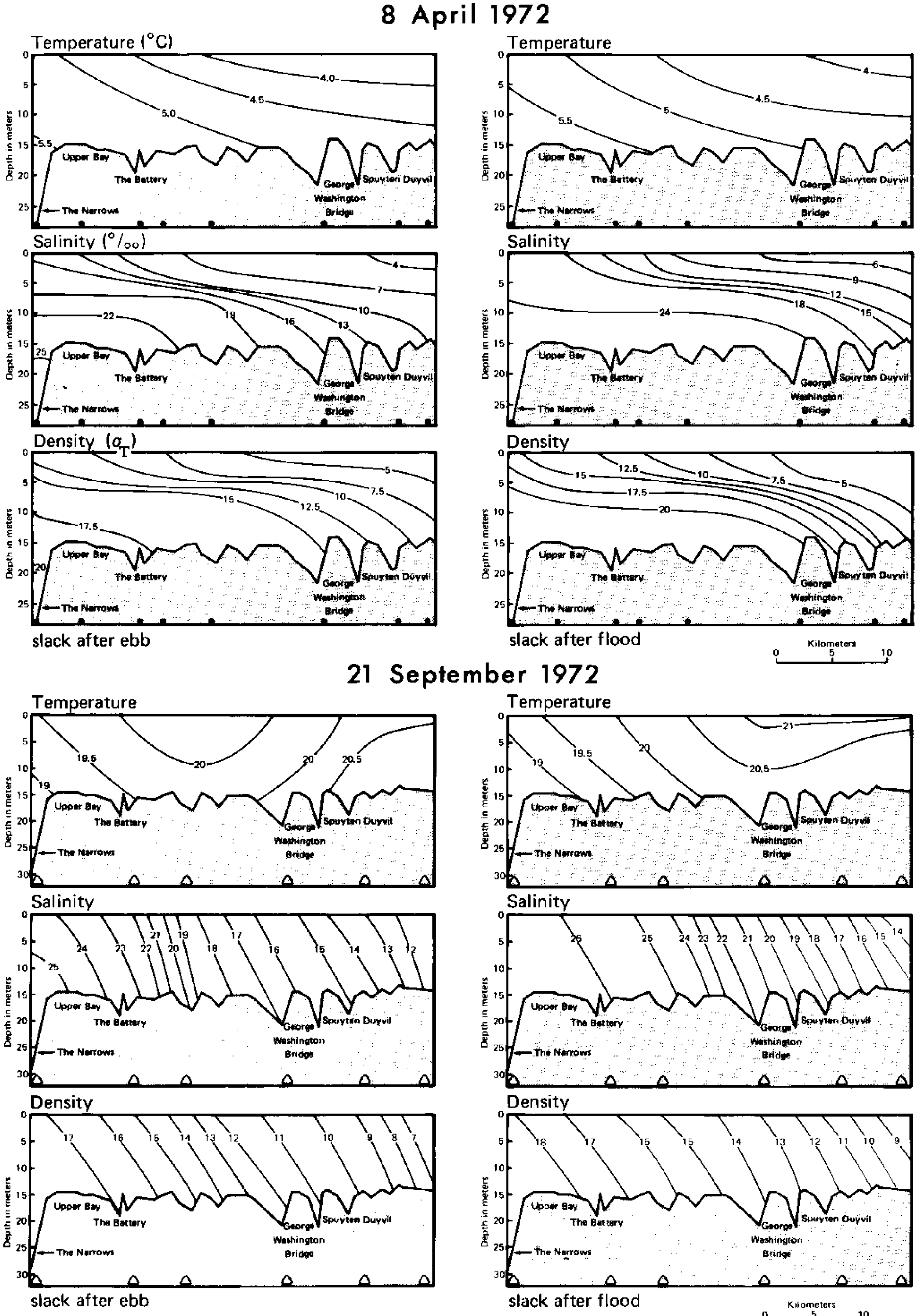


Figure 3. Temperature, salinity, and density contours in lower Hudson River

of the study were 1,450 m^3 sec^{-1} (51,200 cfs) upstream and 3,370 m^3 sec^{-1} (119,000 cfs) downstream, yielding a net seaward flux of 1,920 m^3 sec^{-1} (67,800 cfs). Giese and Barr (1967) estimated the two-month (May and June 1952) mean Hudson River discharge at The Battery at 920 m^3 sec^{-1} (32,500 cfs). To this must be added the discharges of the Raritan, Hackensack, and Passaic rivers (Parsons 1913) and a possible contribution from a net transport through the East River (Jay and Bowman 1975) before Kao's estimates can be compared to Giese and Barr's. Also, fluctuations in discharge throughout the two-month averaging period make it difficult to assess the relative accuracy of the two studies.

Siltation problems in the lower Hudson estuary occur where the river empties into New York Harbor. The abrupt widening causes the salinity structure to change from well mixed to partially mixed. Sediments are transported upstream by the landward bottom flow until they meet the seaward flow in the well-mixed region. Dredging has always been necessary to keep channels and docks in New York Harbor clear of accumulating sediments (Duke 1961; Simmons 1965).

The Hudson Plume

Ketchum, Redfield, and Ayers (1951) made the first extensive survey of the Bight apex during six cruises between February 1948 and January 1950. They described the seasonal distributions of temperature, salinity, density, dissolved oxygen, and iron. Recent studies have been published by Pearce (1970), National Marine Fisheries Service (1972), Charnell and Hansen (1974), and Bowman and Wunderlich (in press).

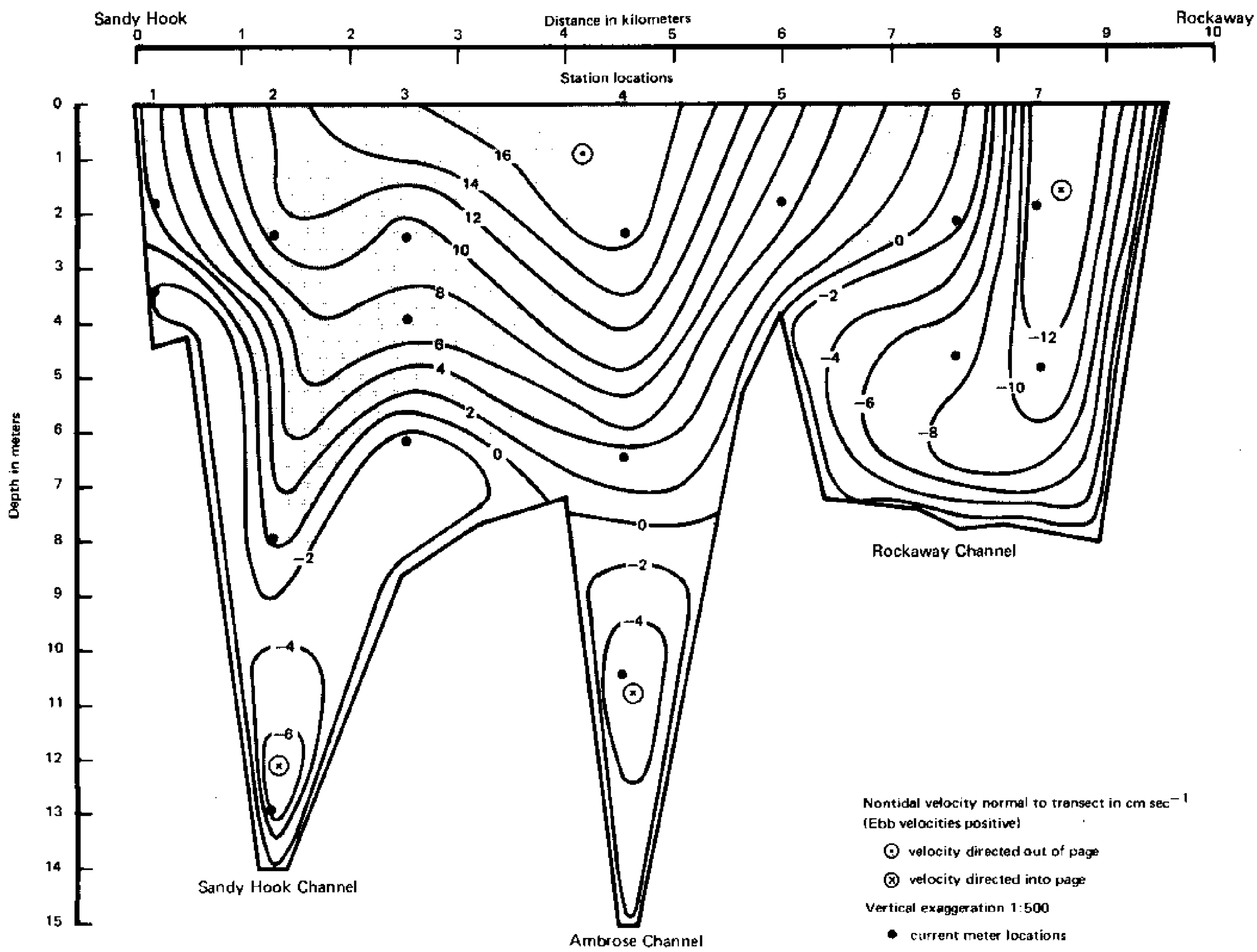


Figure 4. Averaged nontidal velocities, 2-7 June 1952

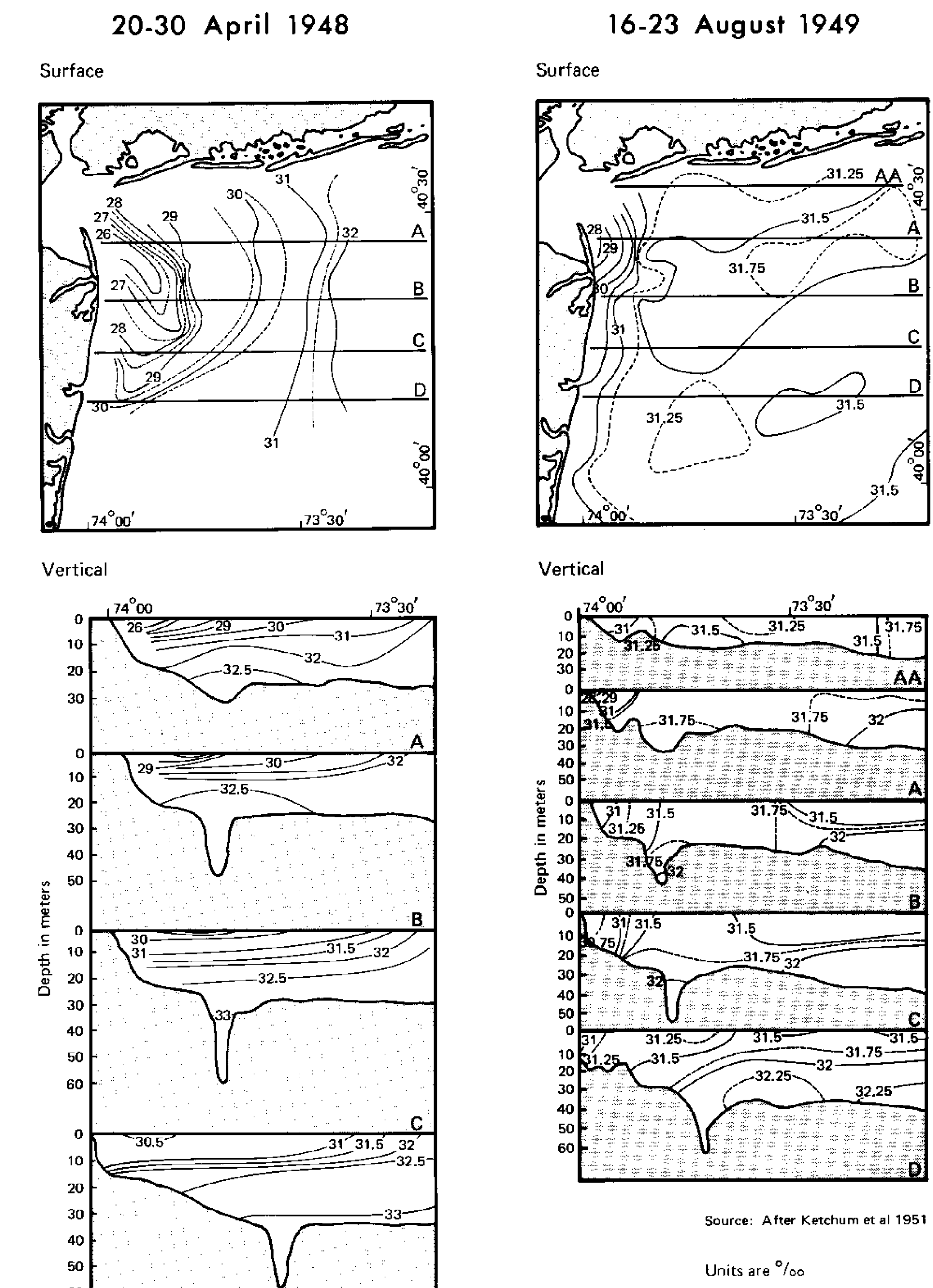


Figure 5. Hudson plume salinity: surface features and vertical sections

Ketchum and his associates showed that each year the total discharges of the Hudson and other rivers entering the Bight displaced a volume of water equal to about 50% of the total volume of the apex. The volume of river water in the area surveyed (about 1,550 km^2) seldom exceeded 1% of the total volume of water. Residence time in the apex for fresh water was estimated at 6 to 10 days, in spite of almost ninefold variations in river flow; it was also concluded that active circulation rapidly disperses the Hudson River *plume*.

During one cruise a late winter storm from the southwest brought about the formation of a well-mixed water column in the vicinity of the Lower Bay entrance. The subsequent reestablishment of the basic circulation pattern of the river plume suggested that only about two days were necessary to recover from the effects of the storm.

Hydrographic conditions in the apex during high discharge (April 1948–estimated runoff about 1,300 m^3 sec^{-1} or 47,000 cfs) and low discharge (August 1949–estimated runoff about 190 m^3 sec^{-1} or 6,600 cfs) are illustrated in the salinity distribution in Figure 5. During high runoff in April the path of the river plume follows the New Jersey shore (Charnell and Hansen 1974); a band of high-salinity water separates the plume from the coast.

Although unambiguous causal relationships are difficult to demonstrate, the set of the plume suggests the combined consequences of two forces: *Coriolis* accelerations deflecting the plume southward and advection by the southwesterly *gradient current* driven by the *steric anomaly* along the coast (Bumpus 1969, 1973). The temperature of the plume during April was not significantly different from the surrounding surface water (6.5° to 8.0°C). East of the axis of Hudson Shelf Valley (the deep channel in vertical sections of Figure 5), the apex water was found to be predominantly shelf/oceanic in character. The April vertical sections indicate the set and depth of the plume (water less than 29°/oo in Figure 5) and also show the strong stratification over the entire apex at this time of year.

Salinity distribution of apex waters during low river discharge in August is also illustated in Figure 5. A weak plume is concentrated around Sandy Hook, although, as is typical in summer, patches of surface water of variable salinity, centered around 31.5°/oo, are spread over the inner shelf. In some areas, vertical salinity stratification exists, whereas in other areas the water is homogeneous. The set of the river plume again lies west of the Hudson Shelf Valley axis.

Surface circulation in the Bight apex is strongly influenced by wind stress. Dominant northward winds during summer contribute to high returns of surface drift cards released into the apex along the New Jersey and southern Long Island shores. In studies by Charnell and Hansen (1974) and Hardy and Baylor (1975) low incidence of drift card stranding in winter was attributed to the seasonally high proportion of winds into the south-to-east quarter.

Studies of circulation near the apex bottom have led to some differing conclusions regarding *residual drift* (Bumpus 1965; Charnell and Hansen 1974; Charnell and Mayer 1975; Hardy et al 1975). Bumpus found an inshore component of the generally along-shore southward drift inside the 60 m contour of the Middle Atlantic Bight. This inshore drift was particularly strong into estuary mouths, including the Hudson estuary.

Charnell and Hansen's investigation of returns of seabed drifters released into the apex also indicated a strong northward flow toward the estuary inside Hudson Shelf Valley; some water diverged eastward along the south shore of Long Island, suggesting the existence of an anticyclonic (clockwise) gyre. Charnell and Mayer found further evidence, from 24 current meter records, of low-frequency anticyclonic circulation during late summer and early winter 1973. The mean flow in portions of the apex consisted of a clockwise circulation of 4 to 10 cm sec^{-1}.

From returns of over 4,000 seabed drifters released into the Bight and apex between January and September 1974, Hardy and his associates found evidence of a persistent drift toward the Hudson estuary mouth. Mean drift velocities were calculated to be about 0.5 to 0.7 cm sec^{-1}. Seabed drifters released in Hudson Shelf Valley consistently drifted northwest in Lower Bay; drifters released from adjacent stations on the valley shoulders had variable trajectories. East of the valley, the bottom drift was northwest to northeastward toward Long Island, and west of the valley the drift was southwest to northwestward toward New Jersey.

Connecticut River

The Connecticut River discharges indirectly into New York Bight via the eastern end of Long Island Sound (Garvine 1975). The river empties a catchment basin of approximately 29,000 km^2, covering Quebec, New

Hampshire, Vermont, Massachusetts, and Connecticut (Meade 1966). The mean discharge is about 560 m^3 sec^{-1} (20,000 cfs), similar to that of the Hudson River; much of the annual discharge is concentrated in April and May.

During the spring peak the flow typically rises to about 2,500 m^3 sec^{-1} or 88,000 cfs (Garvine 1974*b*); surface salinities in eastern Long Island Sound are markedly reduced at this time (Riley 1956).

Near the mouth of the Connecticut River the ratio of the mean river discharge during a tidal cycle to the *tidal prism* is about 2:1 (Ketchum et al 1951); the Hudson estuary ratio is much lower, about 1:100. The reason for the smaller tidal prism in the Connecticut River is that the river's lower reaches are constricted by the bedrock valley in which it flows (Garvine 1974*b*).

Ketchum and Corwin (1964) studied the effect of river runoff on the salinity of shelf waters south of Montauk, Long Island. They found a strong correlation between the mean salinity of continental shelf water to a depth of 60 m south of Montauk and the six-month running mean of the Connecticut River discharge. High river discharge during spring runoff resulted in shelf salinities about 2°/oo less than those from low summer and autumn discharges. Ketchum and Corwin claimed to be able to predict shelf salinity to within 0.3°/oo indicating tight coupling or communication of river water with continental shelf water in this portion of the Bight.

Temperature Distribution

Winter
(Maps 3 and 4—January, February, March)

During winter, water temperatures in the Bight are lowest near the coast and in Lower Bay and increase steadily toward the continental shelf edge. Temperatures drop to their minimum in mid-January, and the minima continue into late February or early March. Temperature contours at given depths roughly follow the coastline.

In the Bight apex, temperatures drop to their annual minimum of less than 2°C. River runoff is low; vertical mixing is strong, leading to an almost unstratified water column, though a completely isothermal apex is seldom found. Bottom temperatures tend to be slightly higher than surface temperatures because vertical mixing cannot keep pace with the rapid cooling at the air-sea interface. The 4°C temperature in Hudson Shelf Valley suggests an intrusion of warm bottom water.

In the greater Bight, bottom temperatures on the shelf range from less than 5°C near shore to about 13°C near the 200 m contour.

Dominant in the winter hydrographic properties of the outer shelf is the existence of a persistent temperature, salinity, and density *front* separating shelf and slope waters. Satellite infrared data show this front to be contiguous throughout the entire Middle Atlantic Bight (Map 1) from as far east as 65°W longitude down to Cape Hatteras. The front usually touches bottom between 75 and 100 m and slopes seaward towards the surface over a horizontal distance of about 25 to 55 km (Beardsley and Flagg 1976). The front forms the upper surface of a temperature and salinity inversion, manifested as a warm, salty intrusion, or tongue, of slope water.

The temperature inversion, seen along the southeastern vertical edge of the isometric block diagrams in Map 4, is persistent from October through May. Although little is known about the dynamics of the formation of this intrusion, it is thought to be a consequence of the sinking of cooled slope surface water. The cooling results from heat lost by evaporation and conduction to the prevailing cold, dry continental polar air mass blowing offshore during winter. These winds also drive a surface flow that transports shelf waters over slope waters, forming a *convergence zone* or front at the boundary of the two water masses. The resulting warm, salty tongue of uniform properties sinks obliquely and intersects the shelf at about 150 to 200 m. This band of slope water has a relatively constant temperature of 9° to 13°C throughout the entire year. The shelf/slope boundary region and its importance in cross-shelf exchange processes is described in detail later.

Spring
(Maps 3 and 4—April, May)

In April the first effects of vernal warming commence most rapidly near the coast where temperatures begin to increase throughout the water column. However, offshore temperatures remain almost unchanged from March and the water column continues to be almost unstratified.

Surface temperatures in the apex reach 7° to 8°C, but bottom temperatures often remain less than 4°C except near Long Island's south shore.

Bottom temperatures on the inner and central shelf in the Bight remain near their annual low of about 5°C. A sharp, horizontal temperature gradient with contrasts of 5°C exists near the shelf break where bottom temperatures are around 10°C.

Strong thermal stratification begins to appear in May as surface waters warm more rapidly than near-bottom waters and the development of the summer *thermocline* begins. Surface temperatures rapidly rise to 9° to 12°C; bottom temperatures remain between 4° and 10°C, forming a "cool pool," a dominant feature of bottom shelf water throughout summer (Bigelow 1933).

Summer
(Maps 3 and 4—June, July, August)

The thermocline that appeared in May continues to strengthen in intensity through June; surface temperatures rise to 17° to 20°C. The temperature front at the shelf break and the associated subsurface, warm intrusion have disappeared as the upper layers continue to warm. Bottom temperatures remain relatively unchanged from May—between 6° and 8°C in Hudson Shelf Valley and on the central shelf. Close to the coastline, however, rapid warming occurs where the depth is less than that of the developing thermocline; temperatures rise as high as 18°C along the New Jersey coast.

The central shelf cold pool of water (Bigelow 1933), a remnant from the previous winter, is usually found between 30 and 100 m and is relatively isothermal both along-shelf and cross-shelf. Temperatures around the pool increase upward, landward, and seaward; the southern extent of the pool penetrates south throughout the Middle Atlantic Bight in August (Boicourt 1973). There is a sharp increase in bottom temperatures seaward of the pool; contrasts of 5°C lie between 90 and 120 m and at 200 m at the continental shelf break.

Bigelow claimed (albeit from widely spaced temperature sections) that since the cold pool was surrounded by warm water, it received no replenishment during the summer from the Gulf of Maine around Nantucket Shoals (Map 1). Ketchum and Corwin (1964) studied the cycle of temperature and salinity in the Bight during 20 cruises between 1956 and 1959. Based on Bigelow's conclusions about the isolation of the cold pool from northern waters, Ketchum and Corwin described the persistence of this cold pool and the annual cycle of temperature on the shelf. By investigating salinity changes in the pool as it warmed from May through September, they were able to distinguish between vertical mixing of bottom shelf waters with low-salinity surface waters and with more saline offshore waters.

However, Beardsley, Boicourt, and Hansen (in press) point out that the prevailing southerly mean gradient current of about 5 cm sec^{-1} throughout the Middle Atlantic Bight (Bumpus 1965, 1969, 1973; Bumpus and Lauzier 1965; Harrison et al 1967; Joseph, Massman, and Norcross 1960) is sufficient to replace the entire water mass on the mid-Atlantic shelf during a six-month period. Entrainment of cold bottom shelf water into the Gulf Stream was observed by Ford, Longard, and Banks (1952) and by Fisher (1972). Stommel (1965) argued that this shelf water must originate near Cape Hatteras. Whitcomb (1970) published data indicating the possibility of a southward transport of cold bottom water around Nantucket Shoals. Unfortunately his spatial resolution is too low to be definitive.

Boicourt (1973) moored five current meters in two vertical arrays near and in the cold pool off Chesapeake Bay over five days in August 1971. Five days was too short to definitively determine the mean velocity fields on the shelf, but a strong southward wind on one day drove a southerly flow that was evident on all current meter records. Whether the cold pool receives replenishment from the north and mixes into the Gulf Stream in the south requires further investigation.

The shelf thermocline continues to intensify during July. Surface temperatures reach their maximum (24° to 26°C) usually early in August, remaining at this peak throughout the month. During these two months, bottom temperatures continue to show distinct rises near shore, due to the intersection of the summer thermocline with the bottom. These isotherms reach higher than 20°C close to shore but only 10°C in Hudson Shelf Valley; they generally follow the Bight bathymetry.

The vertical temperature gradient between the surface and the bottom reaches a seasonal maximum, with contrasts up to 17°C, just before autumnal cooling begins in early September. During summer the water column possesses maximum stability. The depth of the mixed layer is about 10 to 15 m and most of the thermocline is situated between 15 and 30 m.

The thickness of the thermocline increases steadily over the slope where the 15°C isotherm drops from about 25 m to about 50 m. This increase is attributed to greater transparency from low turbidity and productivity in slope waters.

Autumn
(Maps 3 and 4–September, October)

Surface cooling during early autumn begins to break down the summer thermocline at a rate determined by local wind strengths and the roughness of the sea surface (Chase 1959). By the end of October, the isothermal layer has deepened to 20 m in the apex and 40 m in the Bight (Gordon, Amos, and Gerard, in press); surface temperatures have dropped to 14° to 18°C over the shelf. As autumn continues, vertical overturning, associated with heat loss to the atmosphere, and the increase in momentum *flux* from atmosphere to ocean from strong winds reduce vertical temperature gradients. This lowers the average temperature of the water column while increasing the temperature of the bottom water to about 12°C on the central shelf, reducing the volume of the "cool pool." Bottom temperatures along the coast, however, decline during September and October as erosion of the summer thermocline continues.

Early Winter
(Maps 3 and 4–November, December)

Vertical stirring and mixing down to 30 to 50 m occurs during early and middle November. Surface temperatures are between 12° and 17°C over the shelf. During this season shelf temperatures are almost isothermal. For a few weeks the entire water mass on the shelf inside the 80 m contour is almost homogeneous.

During this brief period there is virtually no transition between the bottom waters on the shelf and the isothermal warm zone at the shelf break. Bottom waters on the central shelf reach their seasonal maxima during November and December, sometimes peaking at about 16°C (Ketchum and Corwin 1964).

Most rapid cooling continues closest to the shore through November and December, with temperatures dropping below 8°C due to less heat storage in shallow waters. Isotherms tend to follow the coast. There is a steady horizontal temperature increase at all depths across the shelf, reaching about 12° to 14°C at the shelf break.

Vertical equalization of the water column continues through December, and temperatures decline throughout the Bight. The temperature then drops rapidly to about 6° to 7°C near the coast, within about 2° to 3°C of the winter minimum established in late January. The first indication of the formation of the seasonal subsurface temperature inversion described earlier can be seen in November over the slope at the southeastern corner of the isometric block diagram in Map 4. This tongue intensifies through the winter as the yearly cycle repeats.

Bottom Temperature Maxima and Minima

The isotherms in Map 5 are consistent with the distributions discussed above. However, the contouring is affected by averaging and by the fact that data extremes were utilized rather than the data means used in contouring monthly temperature distributions.

Highest bottom temperatures, exceeding 21°C, are found in the apex. These high temperatures and the sharp horizontal gradients seaward are attributed to the intersection of the summer thermocline with the shelf bottom. A large pool of warm water with temperatures between 15° and 16°C, which lies over much of the central shelf, forms during overturning in November when the water column is well mixed.

Maximum bottom temperatures seaward of this broad band of relatively isothermal water drop rapidly to about 12°C along the 200 m contour (Map 5) and are usually observed in late summer and autumn.

Coldest winter temperatures (less than 1°C) are found closest to the shoreline, especially along the coast of New Jersey, during January, February, and March. Temperatures increase steadily across the central shelf where they range from 2° to 5°C during March and April. Late winter temperatures increase rapidly near the shelf break to over 7°C at the 200 m contour. Hydrographic properties near the continental shelf edge are affected by aperiodic intrusions of cold, salty bottom slope water, and the associated disappearance of the subsurface temperature/salinity maxima.

Discussion

Temperatures in New York Bight have a seasonal cycle so well defined that averaging many years of data has not blurred general characteristics, as a comparison of averaged data with those from individual cruises will show. The main features of the temperature field are: the vertically isothermal structure in winter—coldest near shore but increasing steadily seaward to the slope or oceanic front; the winter temperature maximum between 100 and 200 m under the front; the development and decay of a strong summer thermocline; and the summer pool of cold bottom shelf water formed during the previous winter.

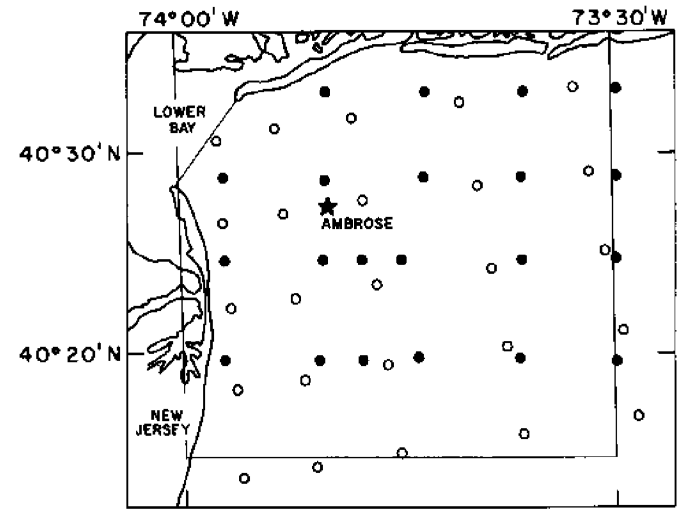
The existence of the shelf break warm zone has been known for almost 100 years (Bigelow 1915). In 1882 Verrill found the bottom shelf water decidedly colder than the previous year and no trace was found of the warm belt between 150 and 200 m. Associated with the disappearance of the warm belt was the extraordinary mortality, by the millions, of the tilefish (*Lopholatilus chamaeleonticeps*), which inhabits this relatively warm water mass. Reduced almost to extinction, tilefish gradually increased again—in 1917, for example, about 11.5 million lb (4.3 million kg) were taken commercially (Bigelow and Schroeder 1953; McHugh 1972).

Measurements obtained by Volkman and Moore (1972) at a fixed hydrographic station (39°10'N, 70°00'W) at the outer edge of the continental slope have shown the occasional intrusion of cold *meanders* from the Labrador current. These intrusions may account for aperiodic disappearances of the inversion.

The Bight apex hydrography is in many ways similar to that near the mouths of other large rivers in drowned river valleys—for example, Delaware and Chesapeake bays (Joseph et al 1960; Harrison et al 1967; Walford and Wicklund 1968; Boicourt 1973; Boicourt and Hacker 1976). However, the bend of the Bight coastline, the deep Hudson Shelf Valley, and seasonal and short-term wind patterns significantly influence the nearshore circulation.

Temperatures in the apex closely follow those of other nearshore regions of the Middle Atlantic Bight. There is a large range of about 25°C between summer and winter surface temperatures. As in other shallow areas of the Bight, strong bottom gradients persist through the summer over most of the apex as a consequence of the intersection of the seasonal thermocline with the bottom.

Map 3. Temperature distribution, Bight apex, 1969 and 1973



Map to left shows station locations

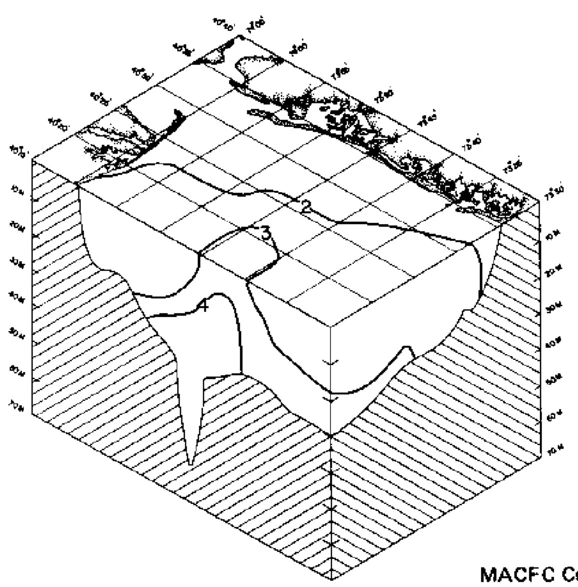
- ○ MESA 1973-1974
- ● Middle Atlantic Coastal Fisheries Center (MACFC) 1969

Map pairs show surface and vertical contours in block diagram, bottom contours in map

Units are °C

Data Sources: Middle Atlantic Coastal Fisheries Center 1972; Hazelworth et al 1974; Hazelworth and Weiselberg 1975

6-7 FEBRUARY 1969



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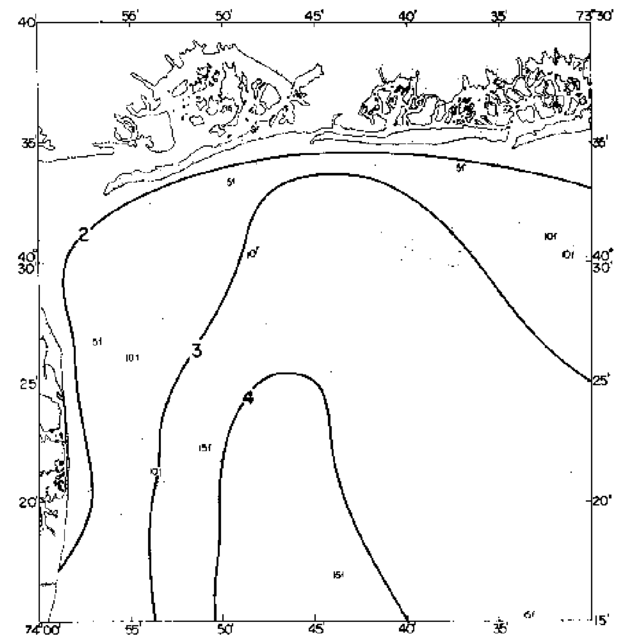
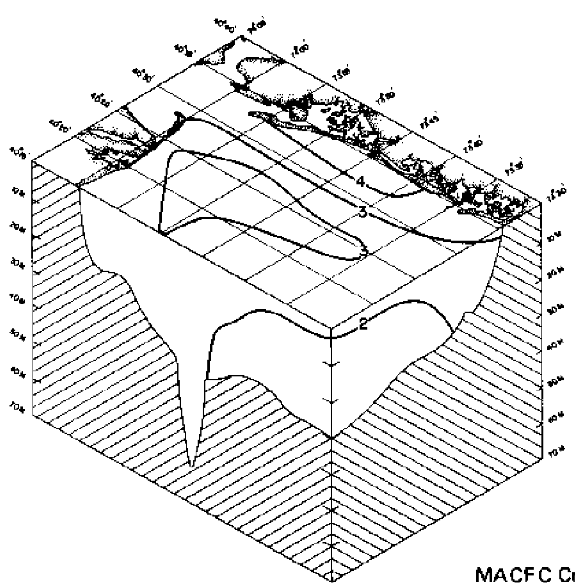
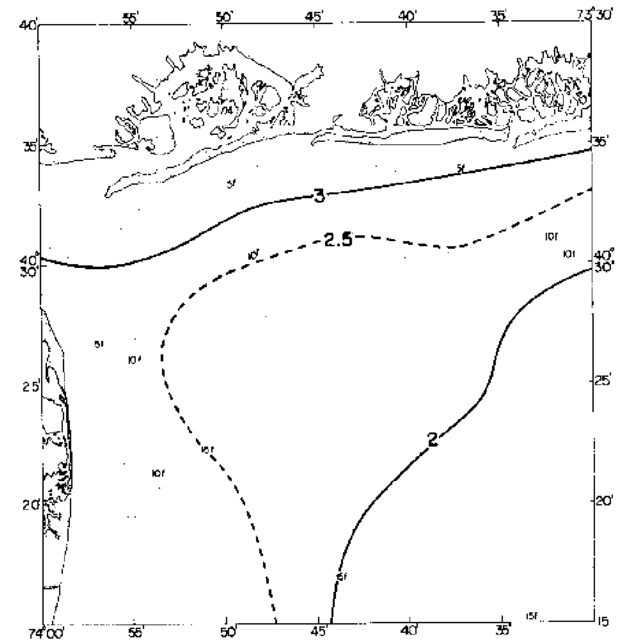


18-19 MARCH 1969



MACFC Cruise 2



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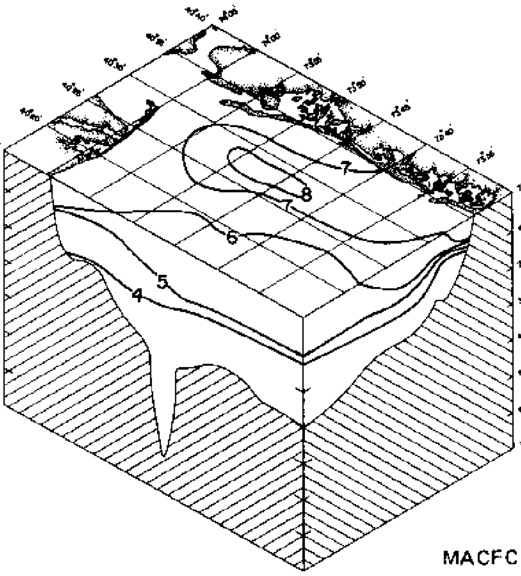
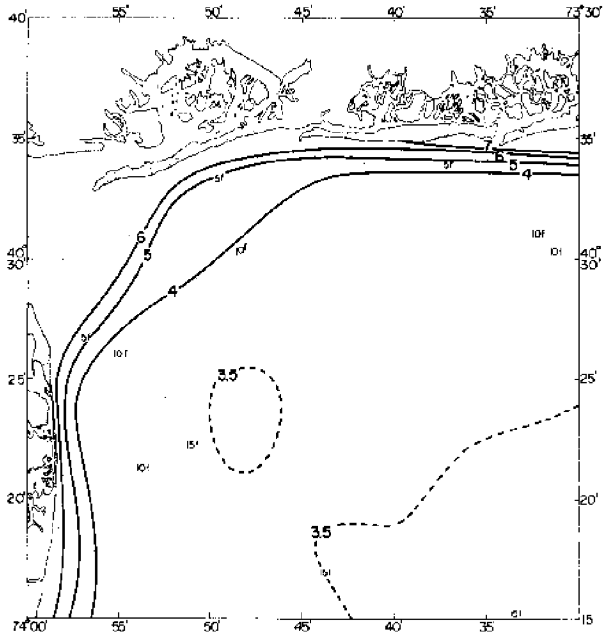


MACFC Cruise 3



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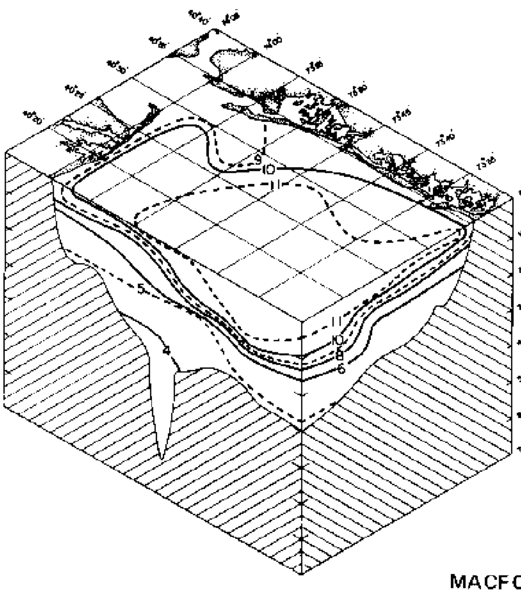
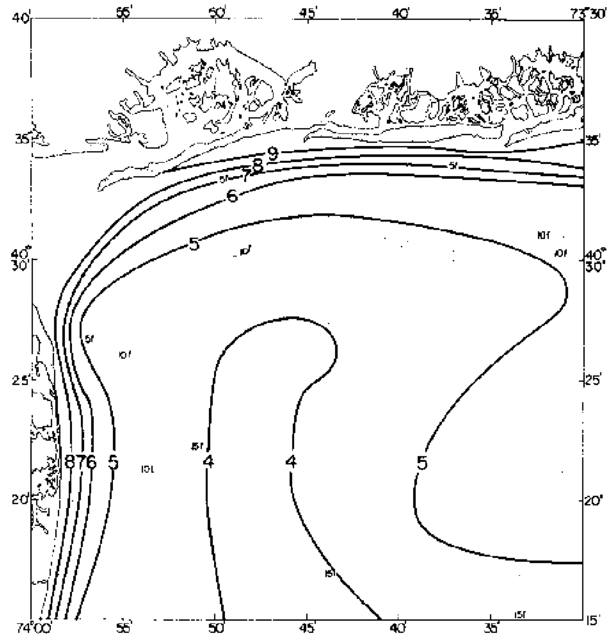


MACFC Cruise 5



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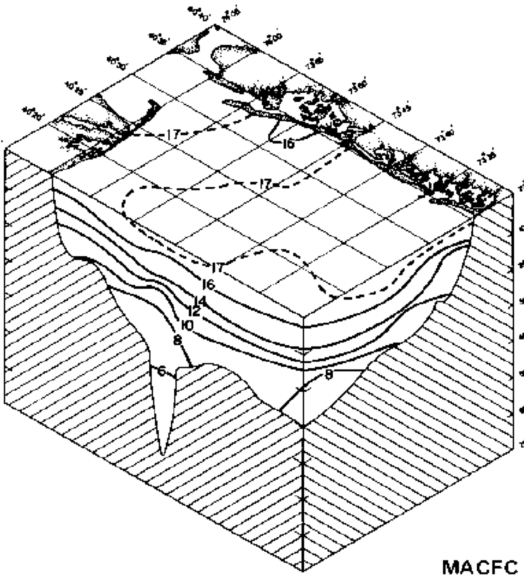

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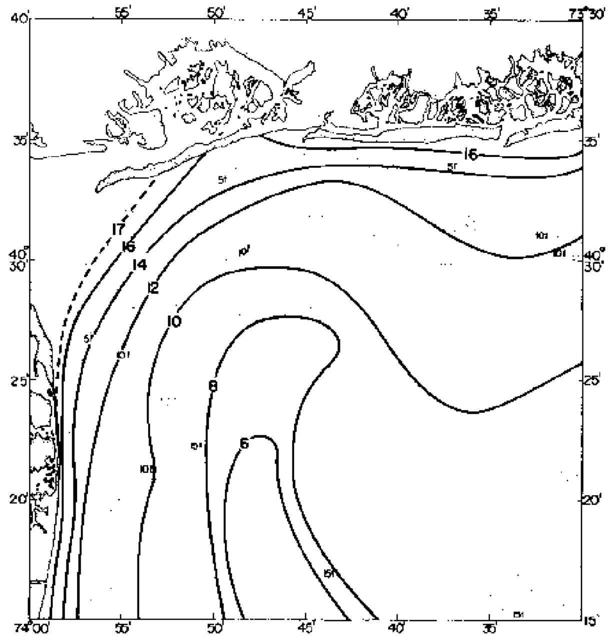

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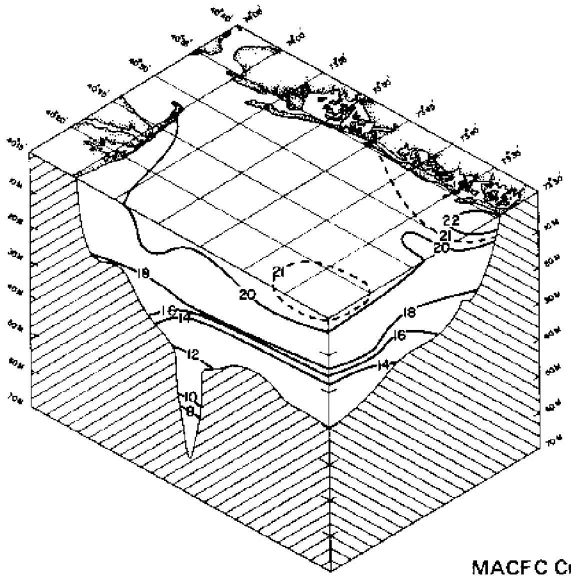

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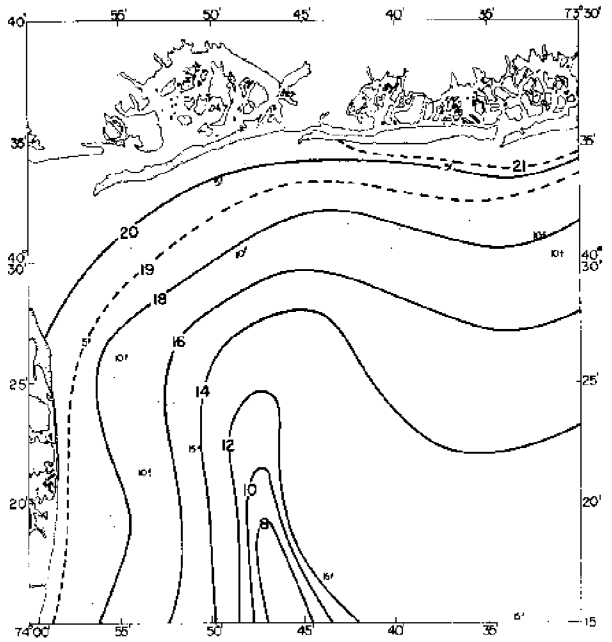
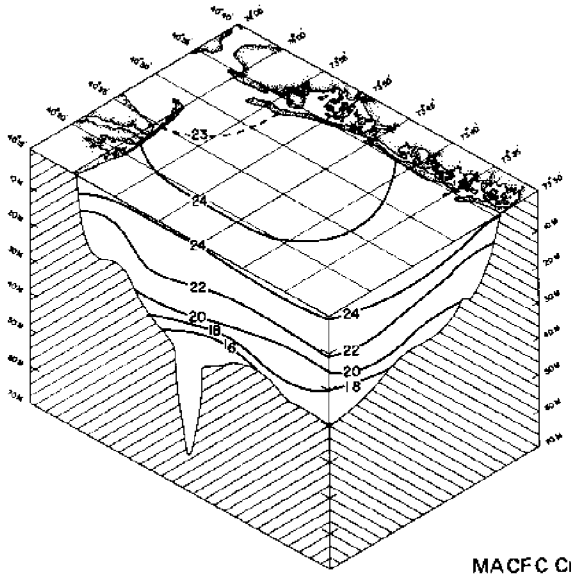


28 JULY-1 AUGUST 1969



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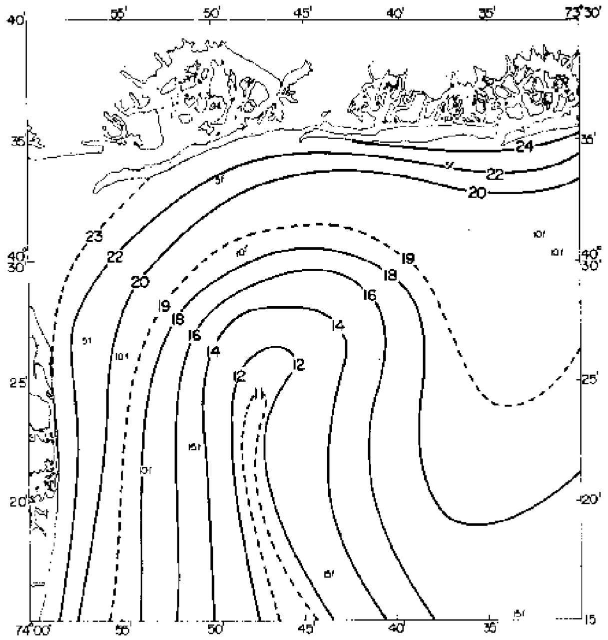

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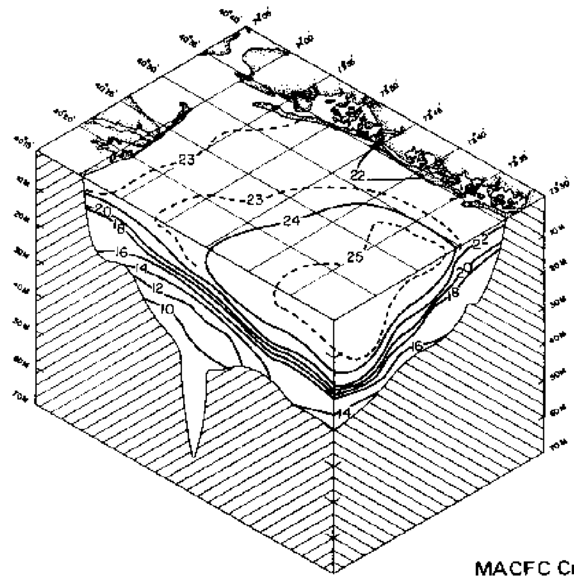
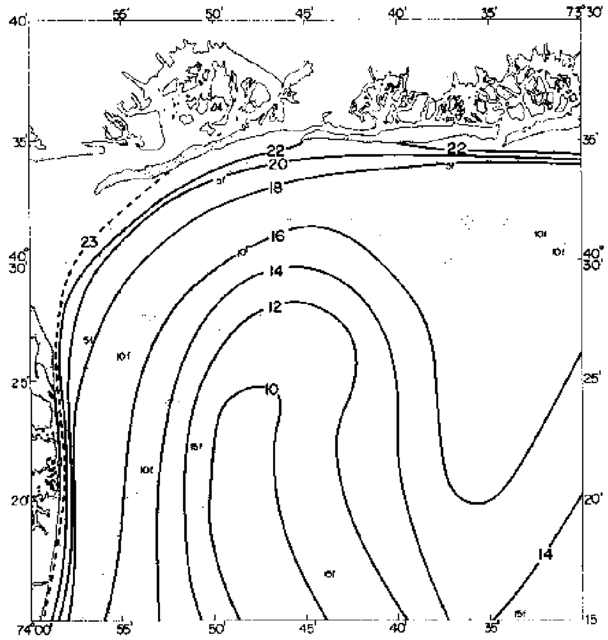


MACFC Cruise 9



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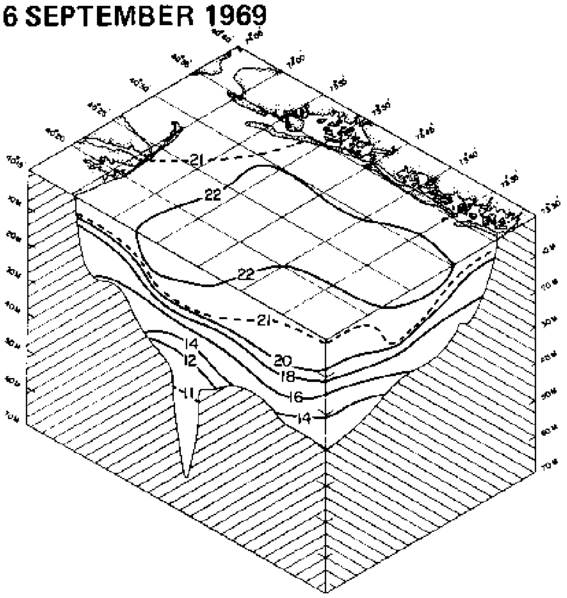
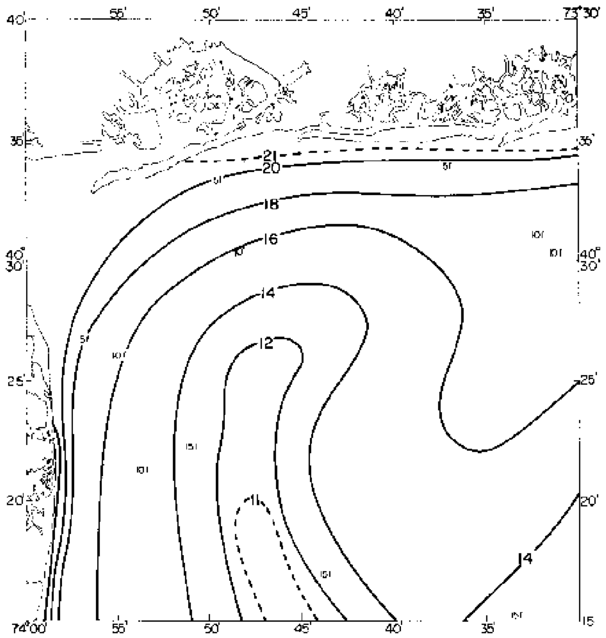
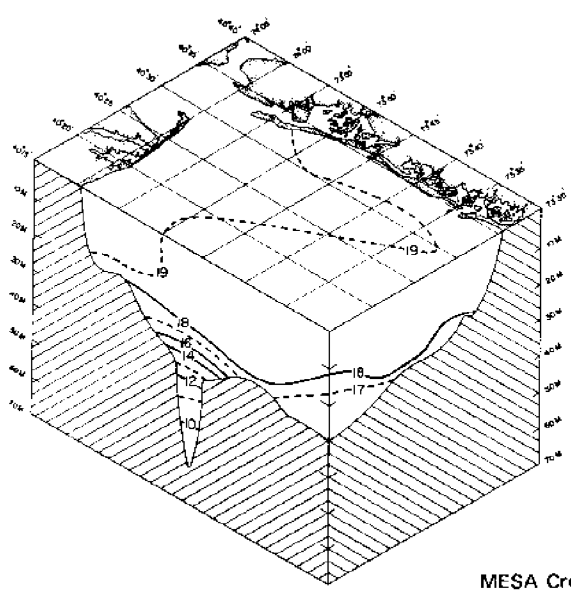
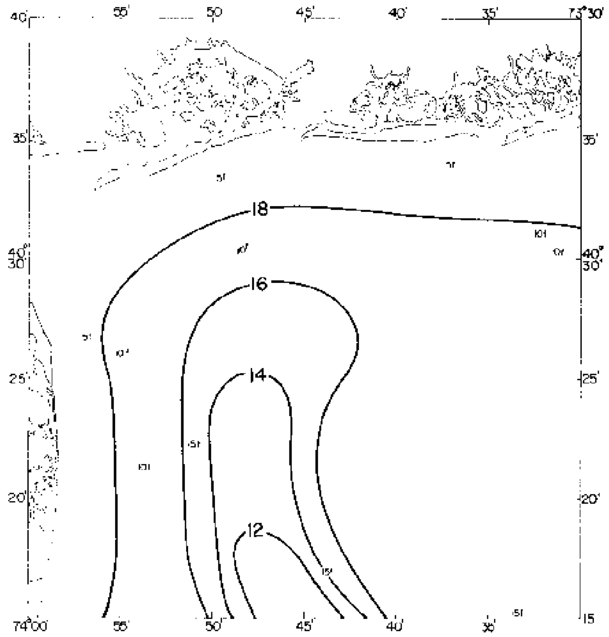




MACFC Cruise 10

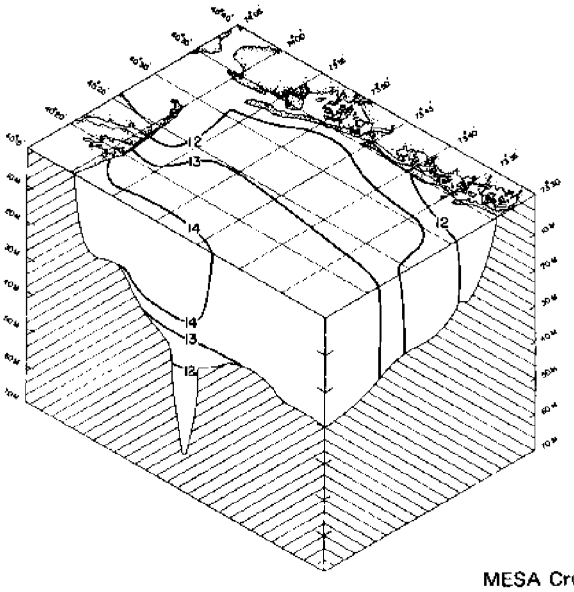
1-4 OCTOBER 1973

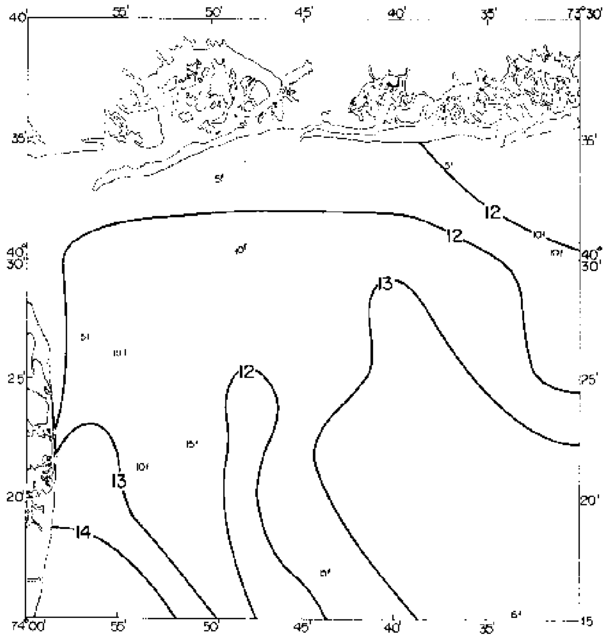




MESA Cruise 3

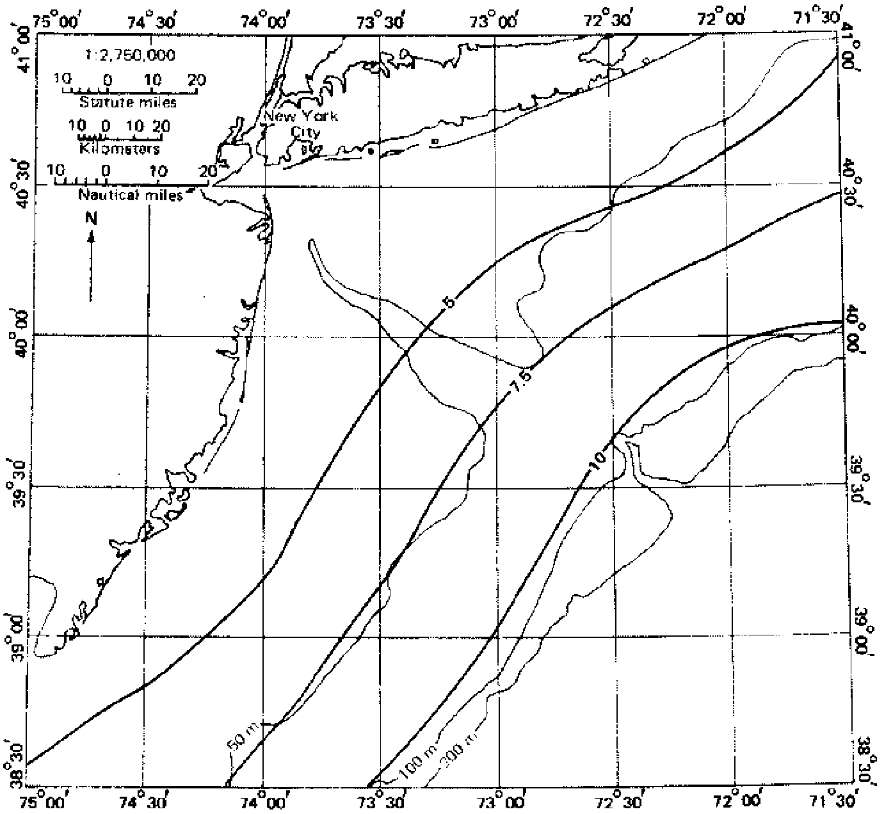
5-9 NOVEMBER 1973





MESA Cruise 4

Map 4. Mean temperature distribution in Bight by month



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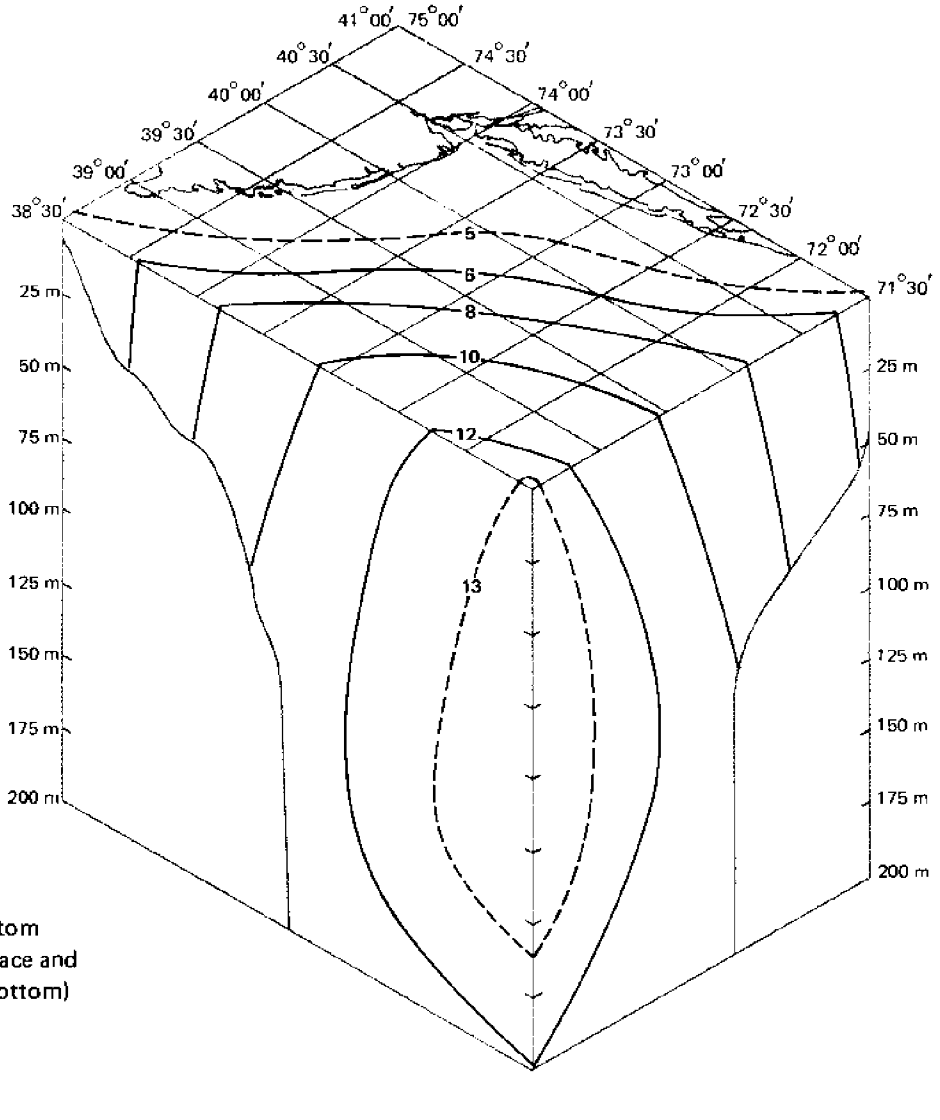
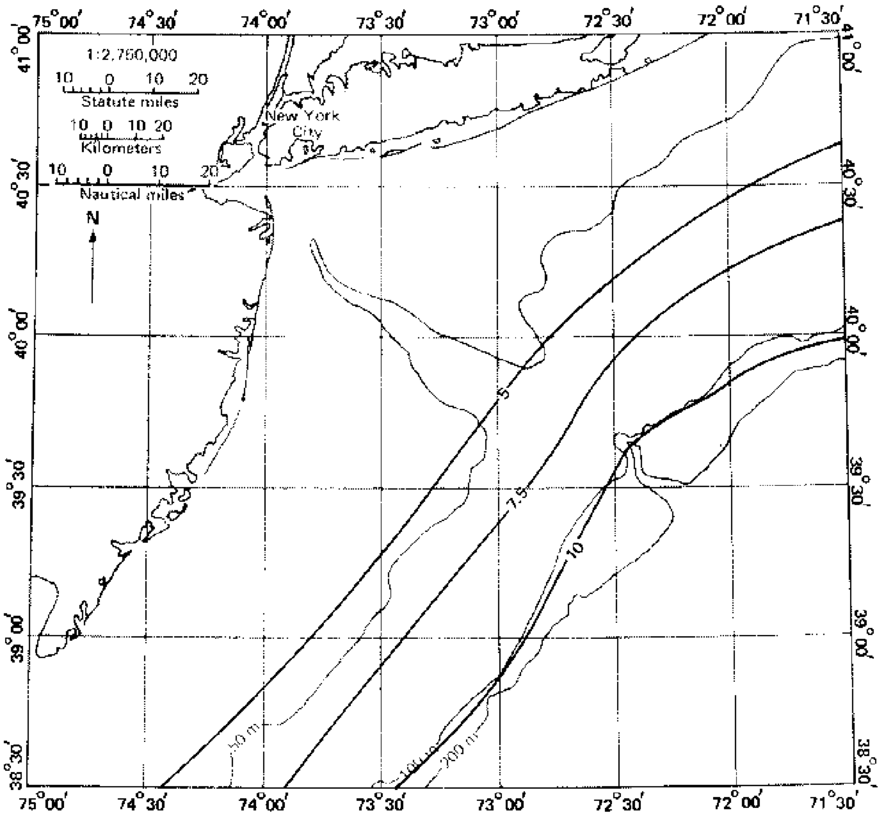


Map pairs show bottom contours (top), surface and vertical contours (bottom)

Units are °C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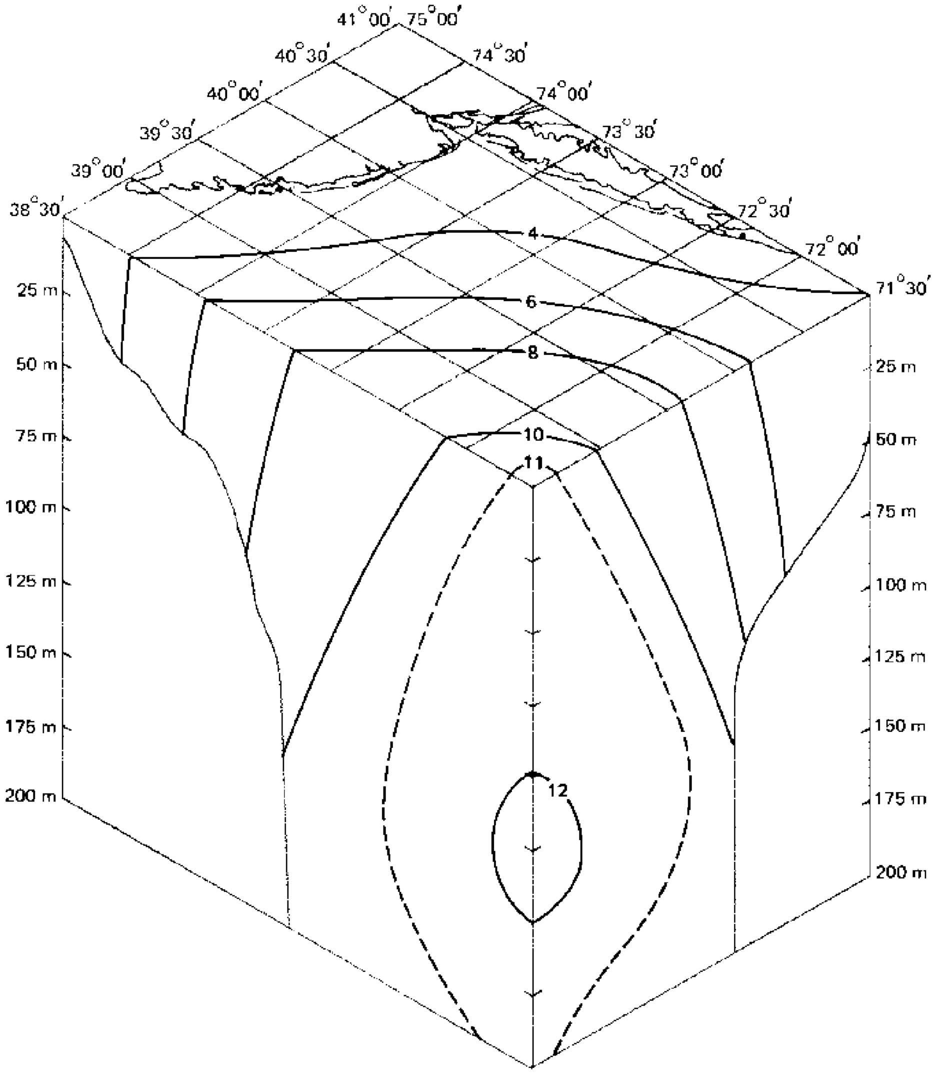
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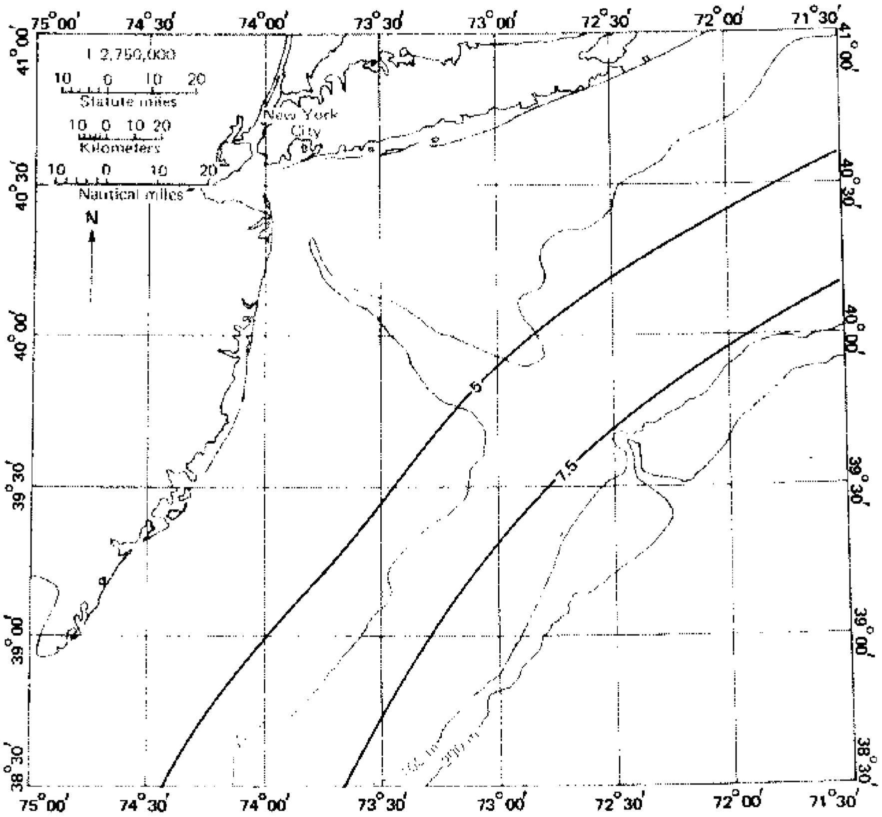
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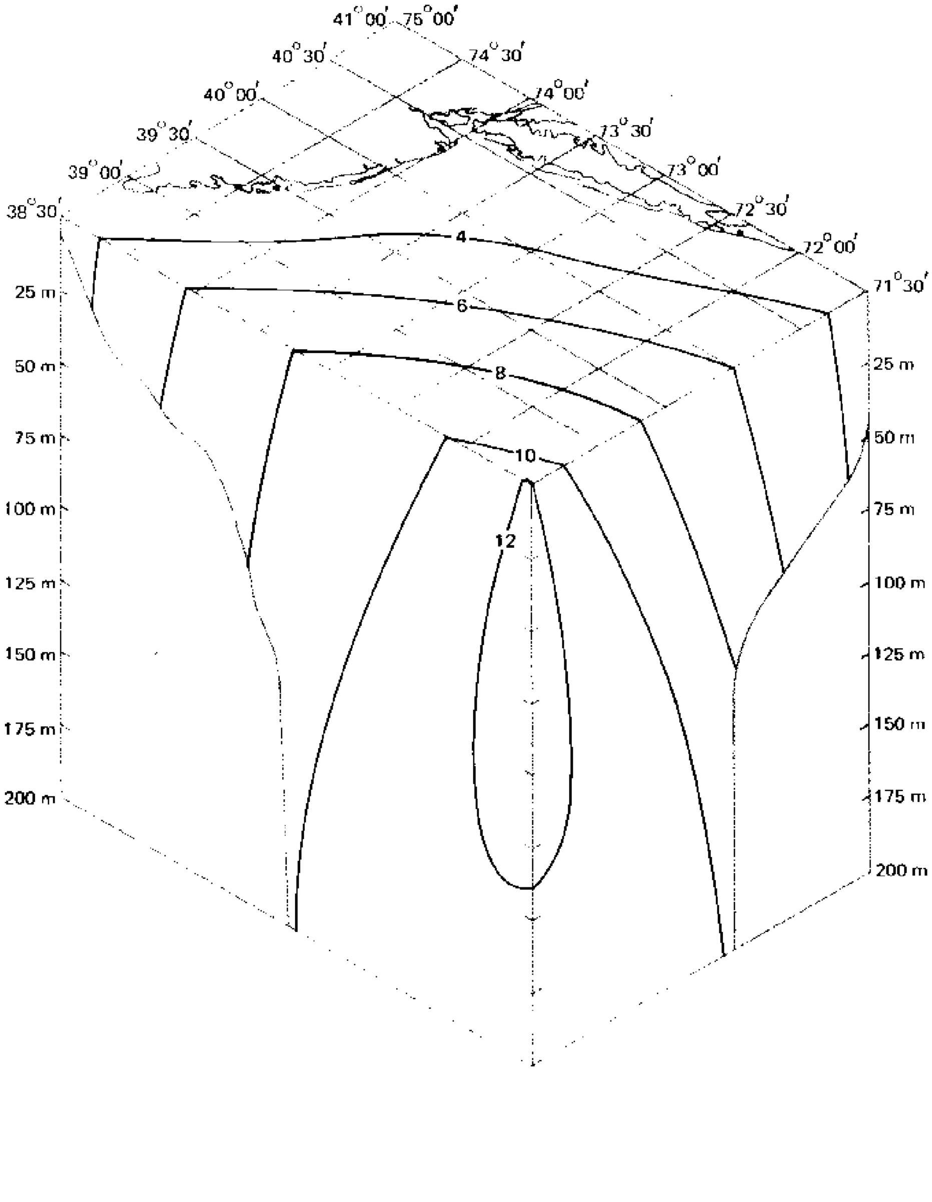
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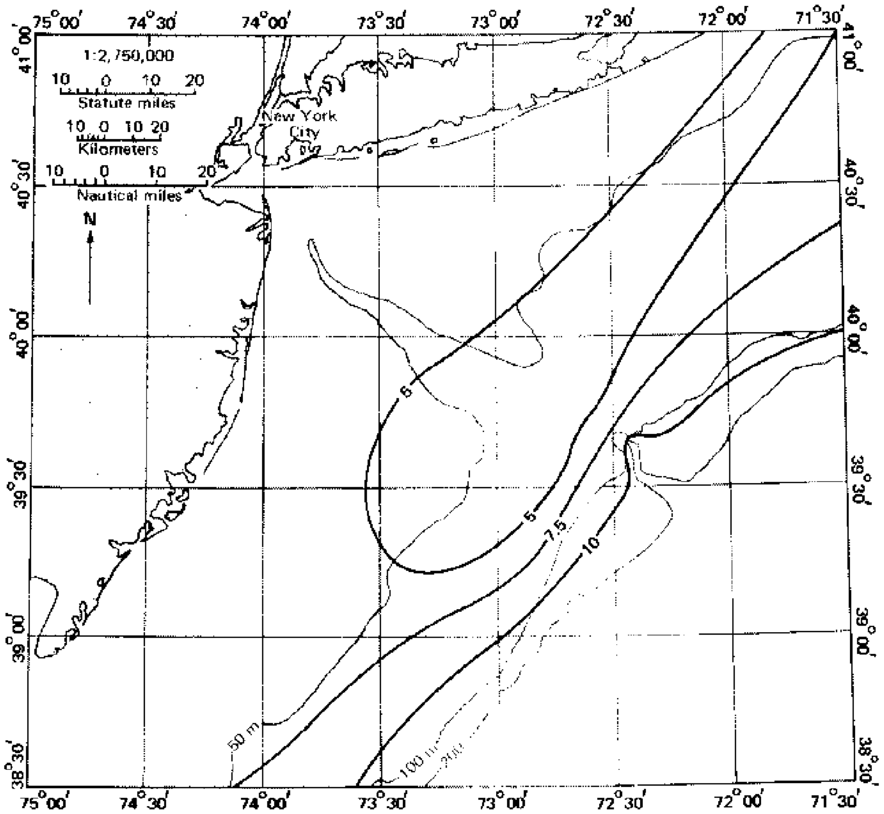
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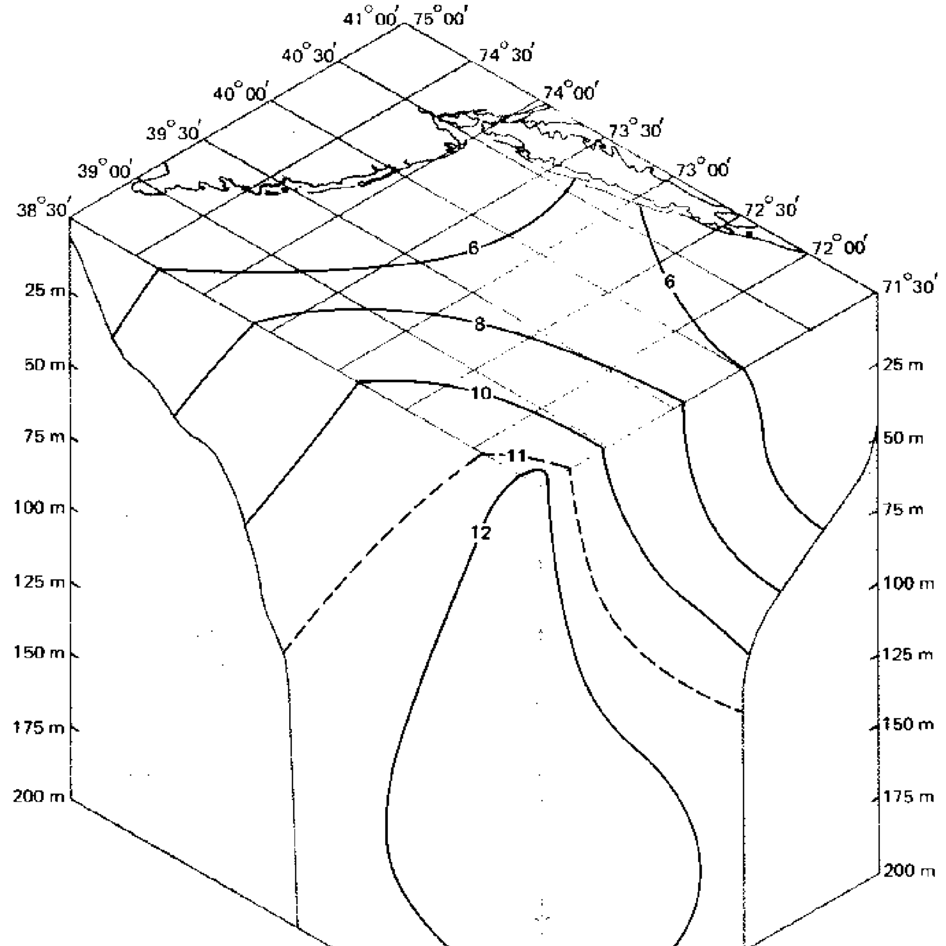
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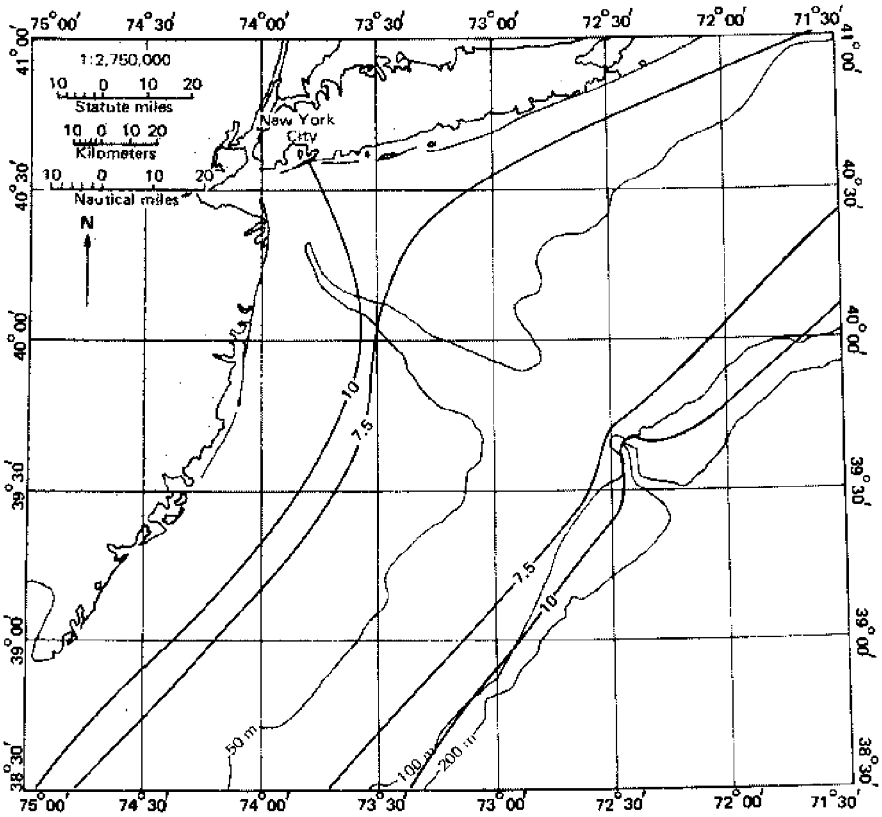
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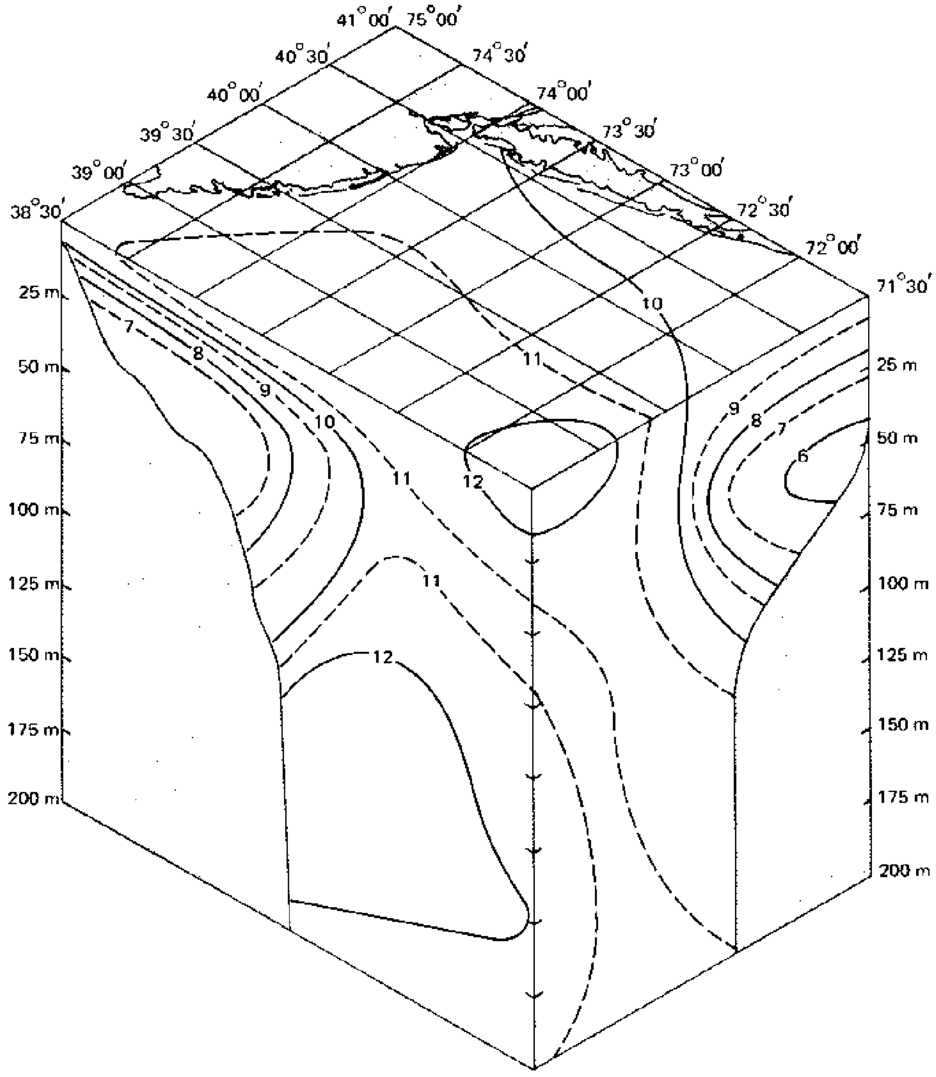
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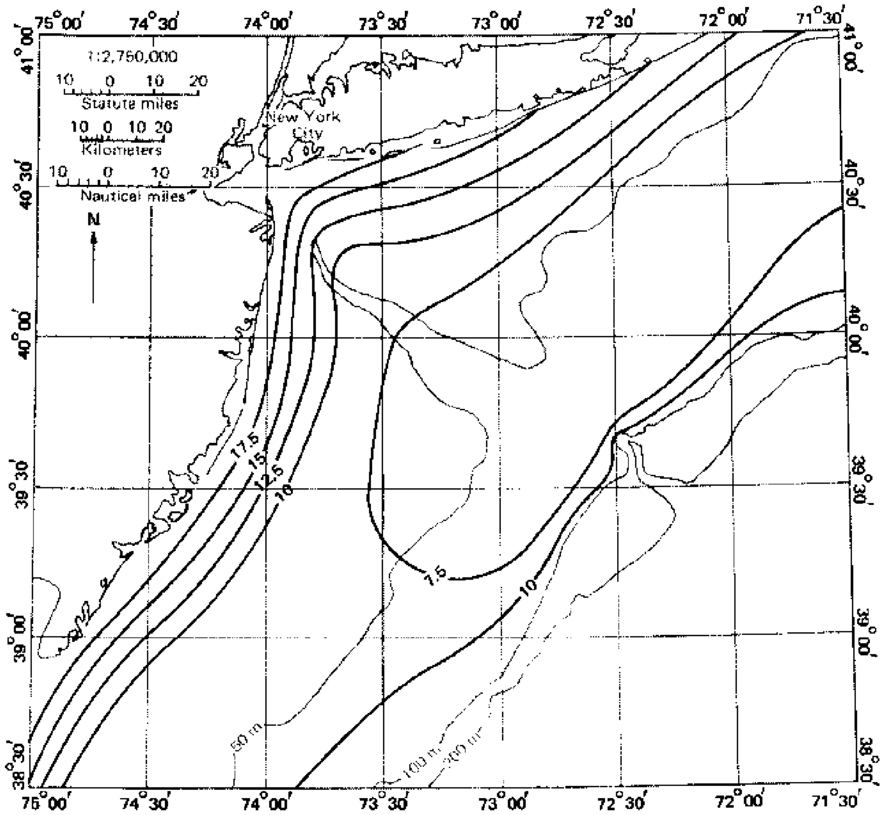
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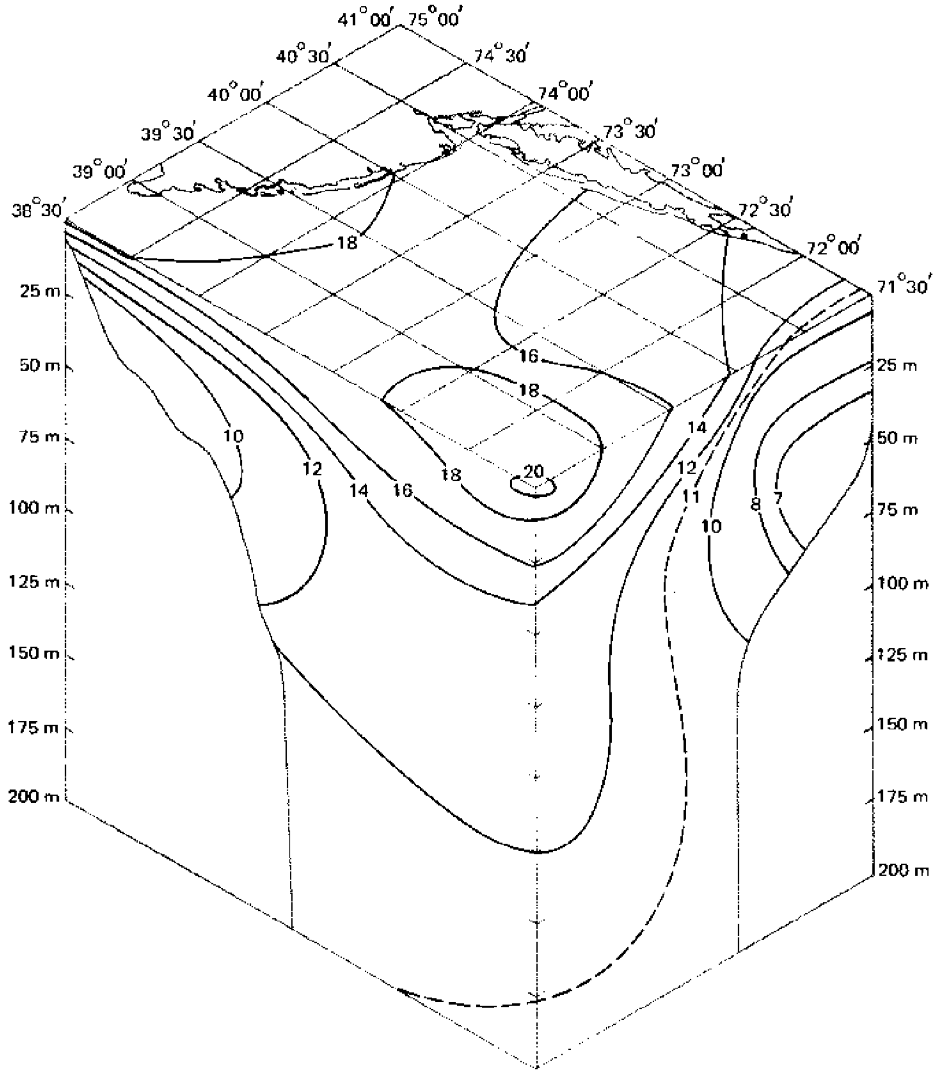
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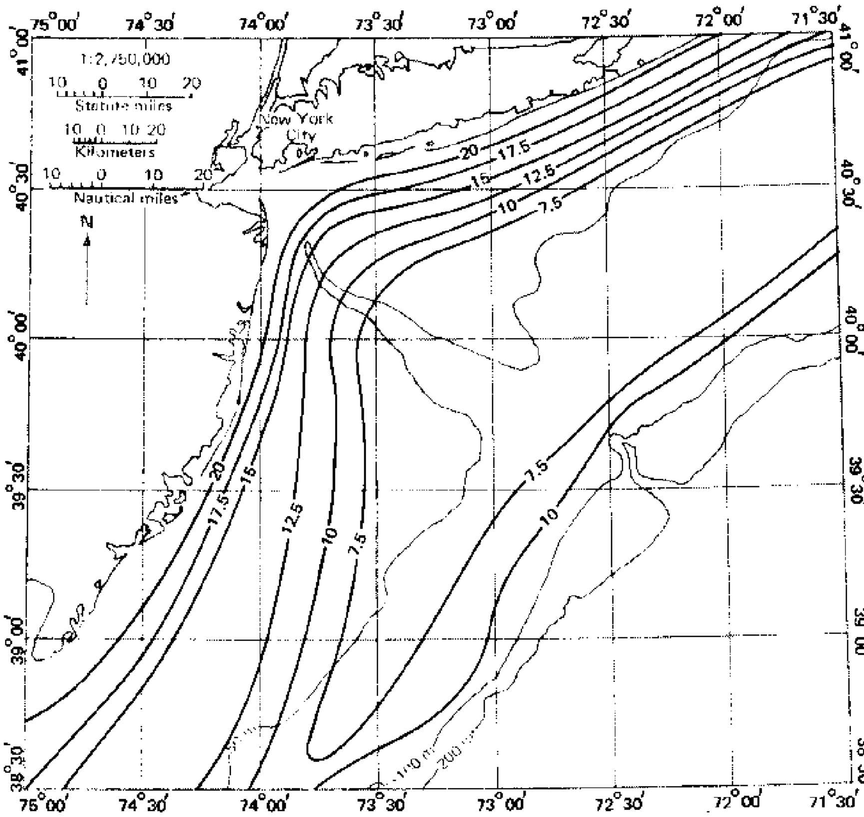
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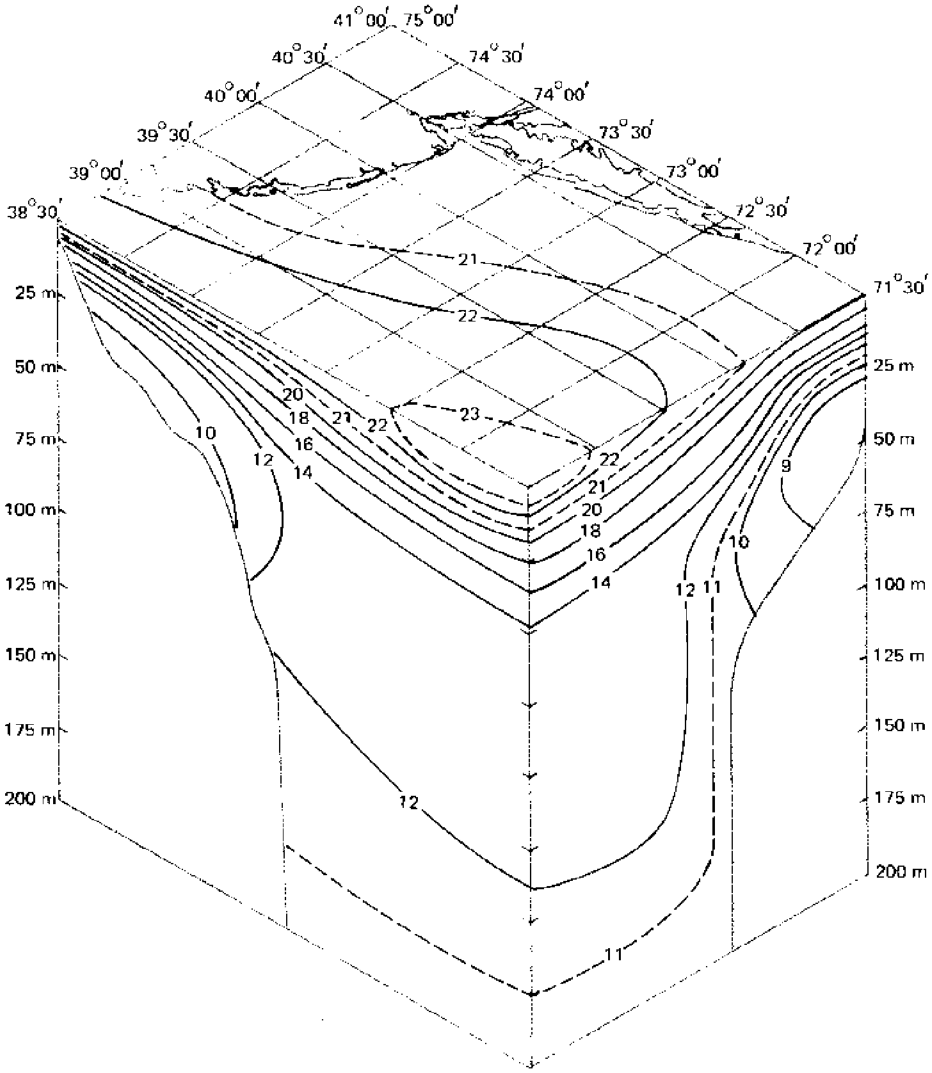
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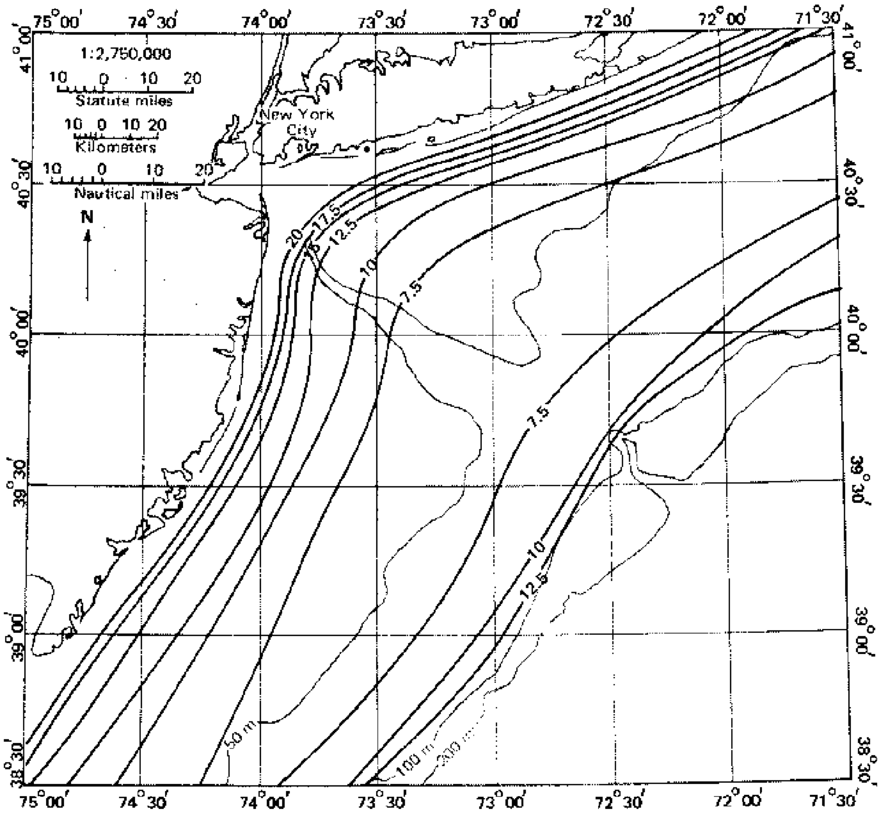
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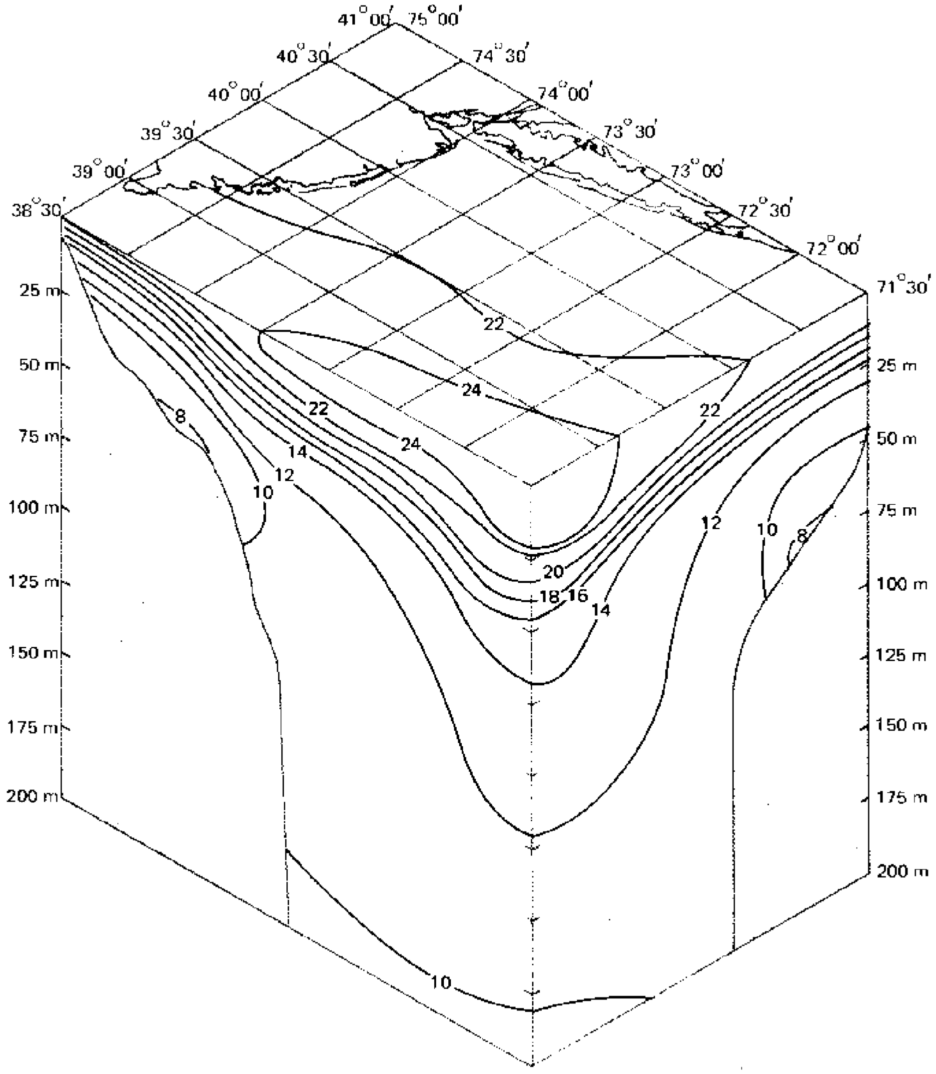
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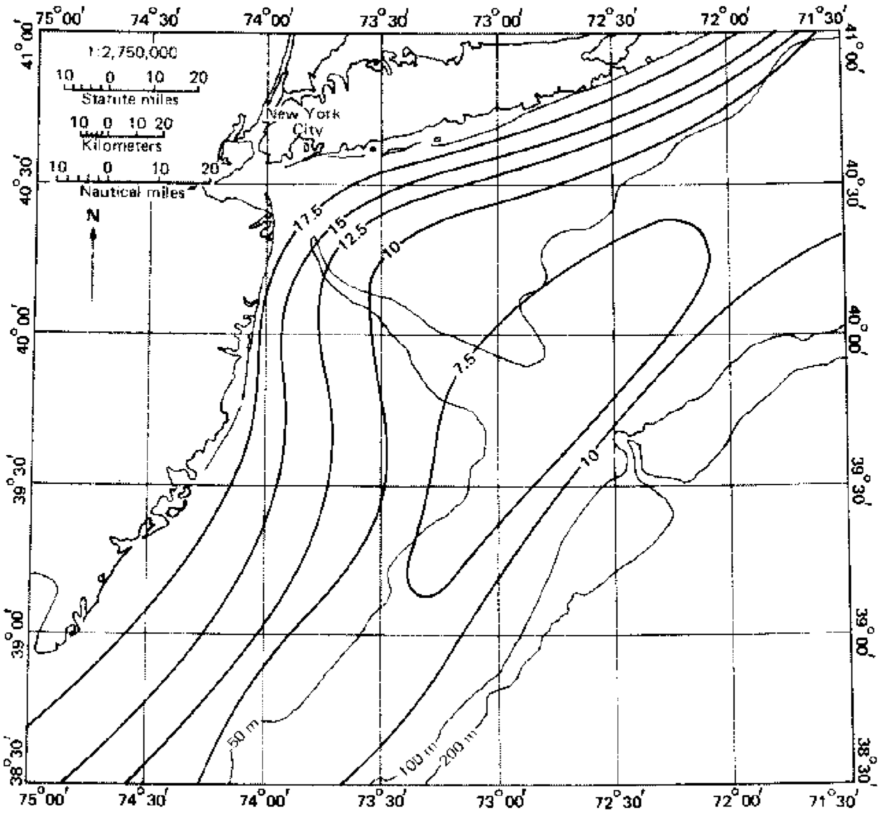
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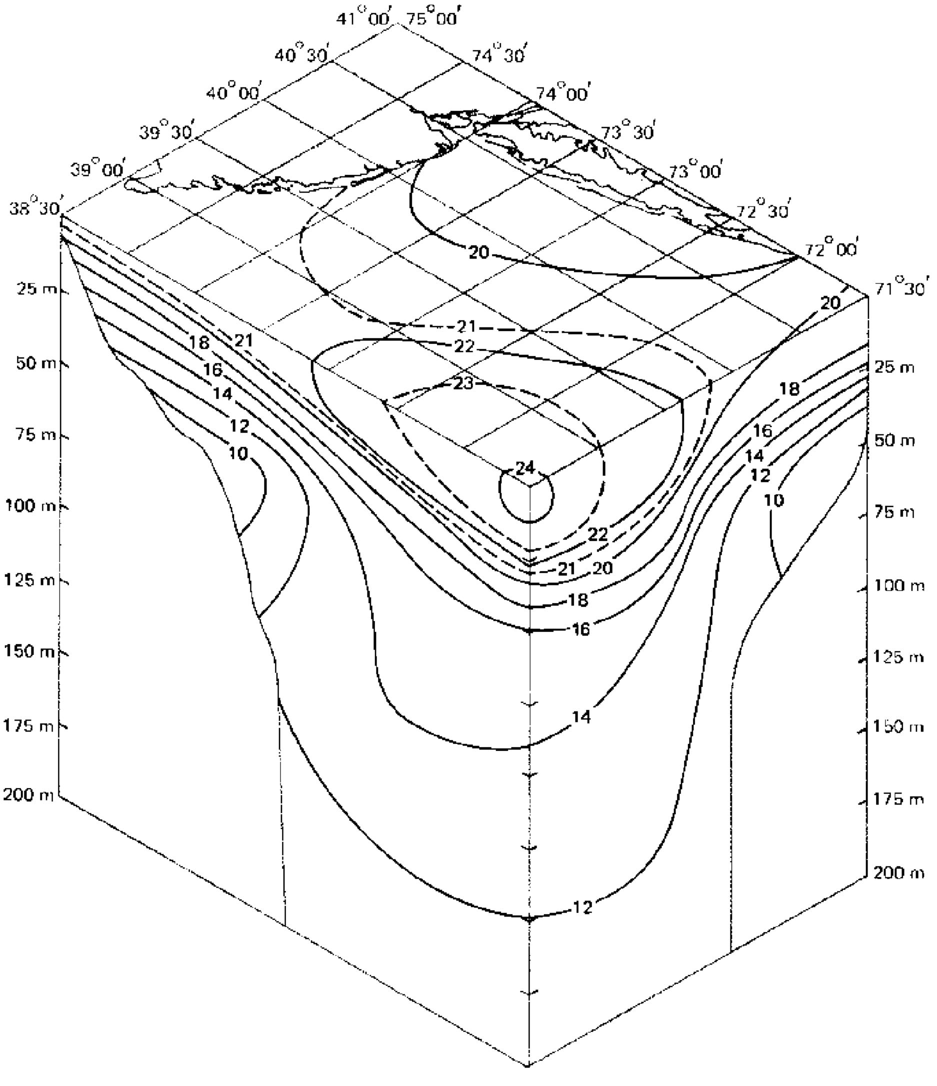
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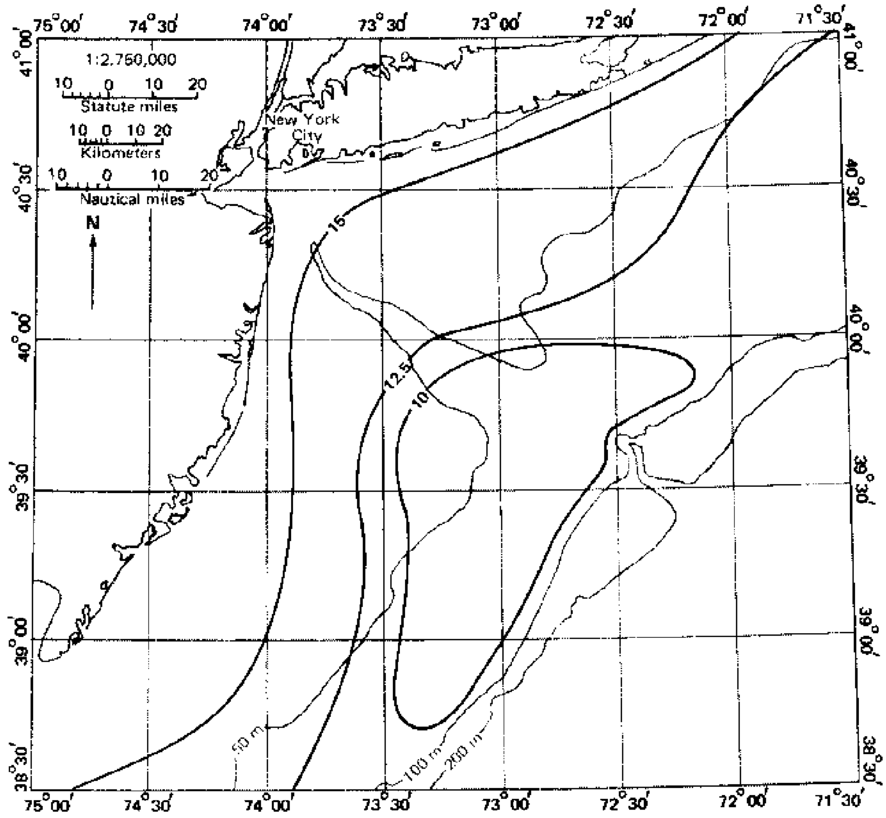
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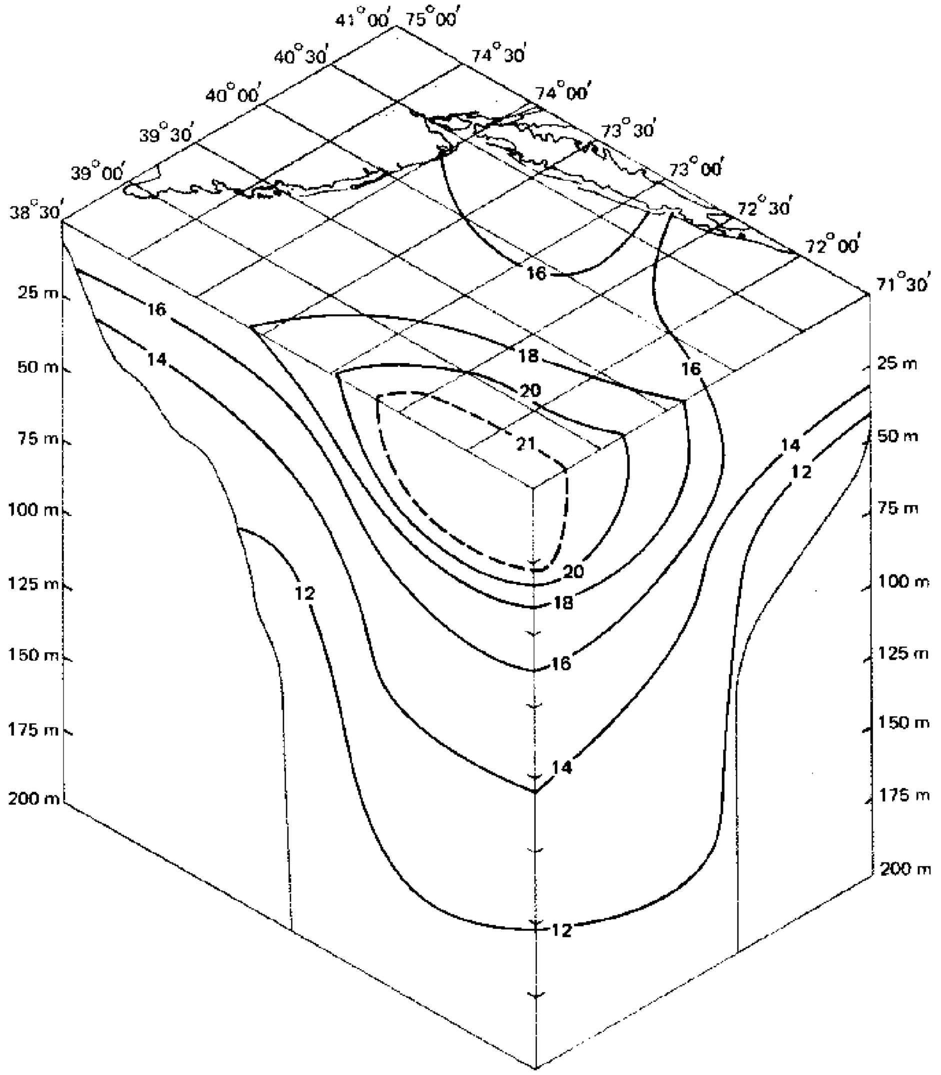
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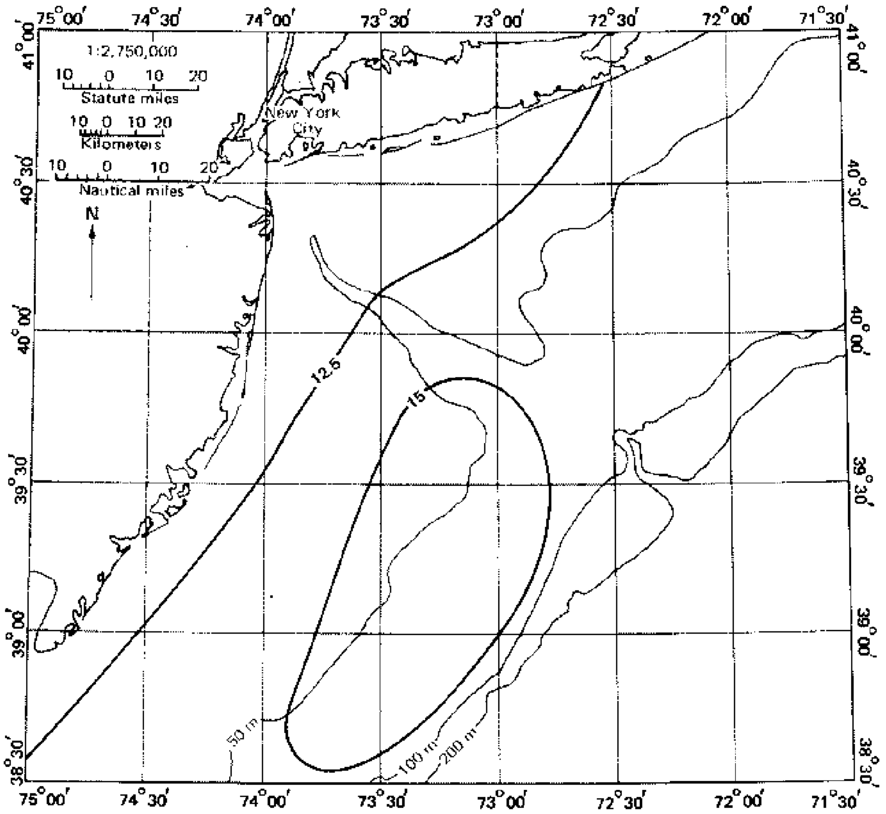
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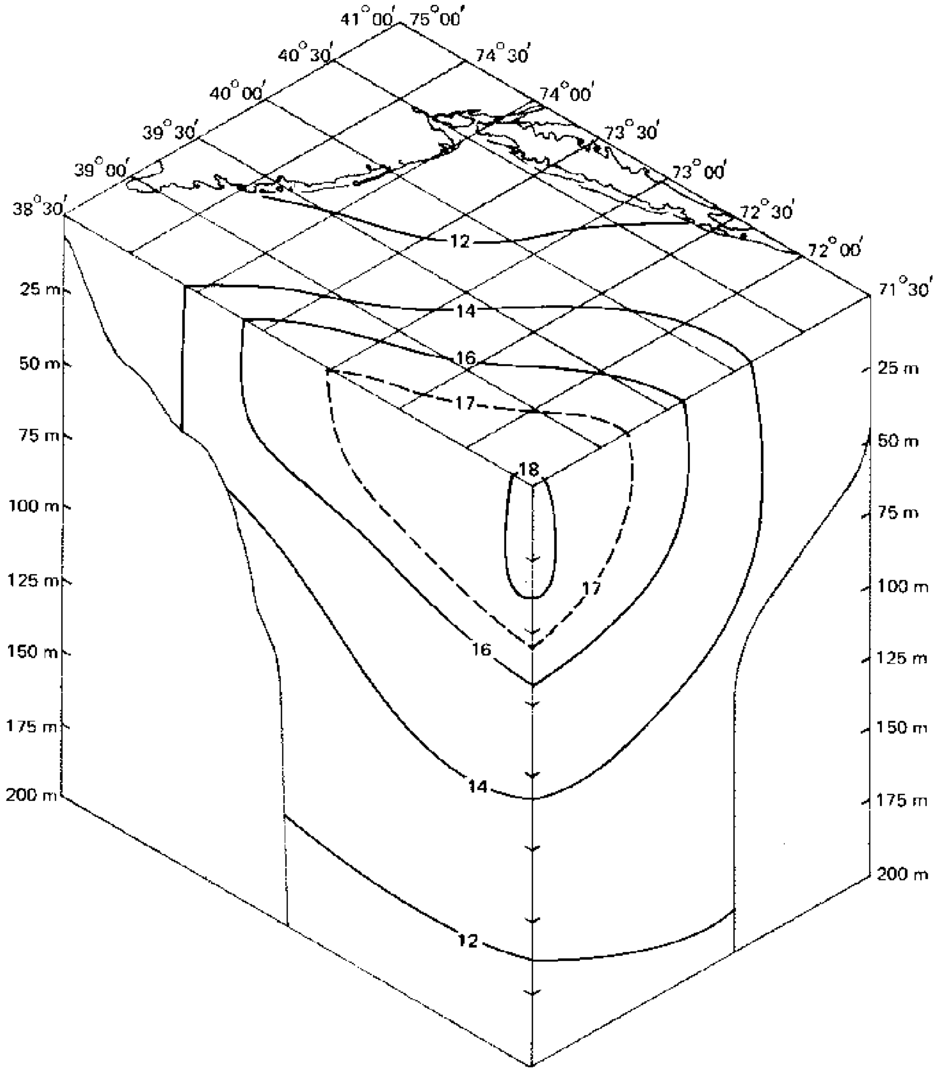
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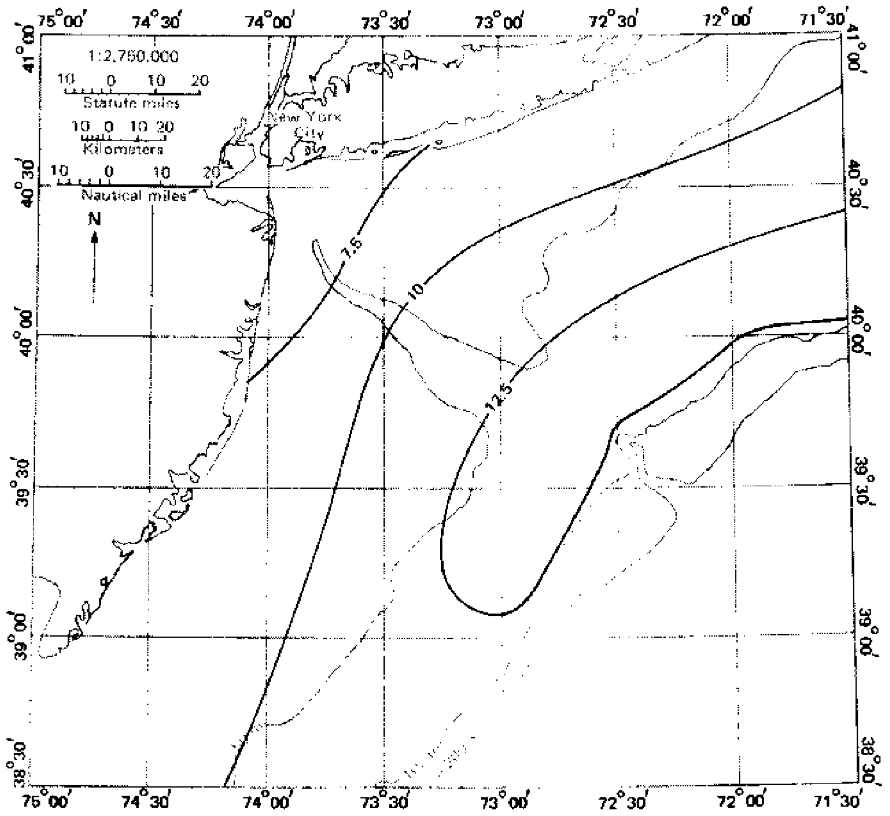
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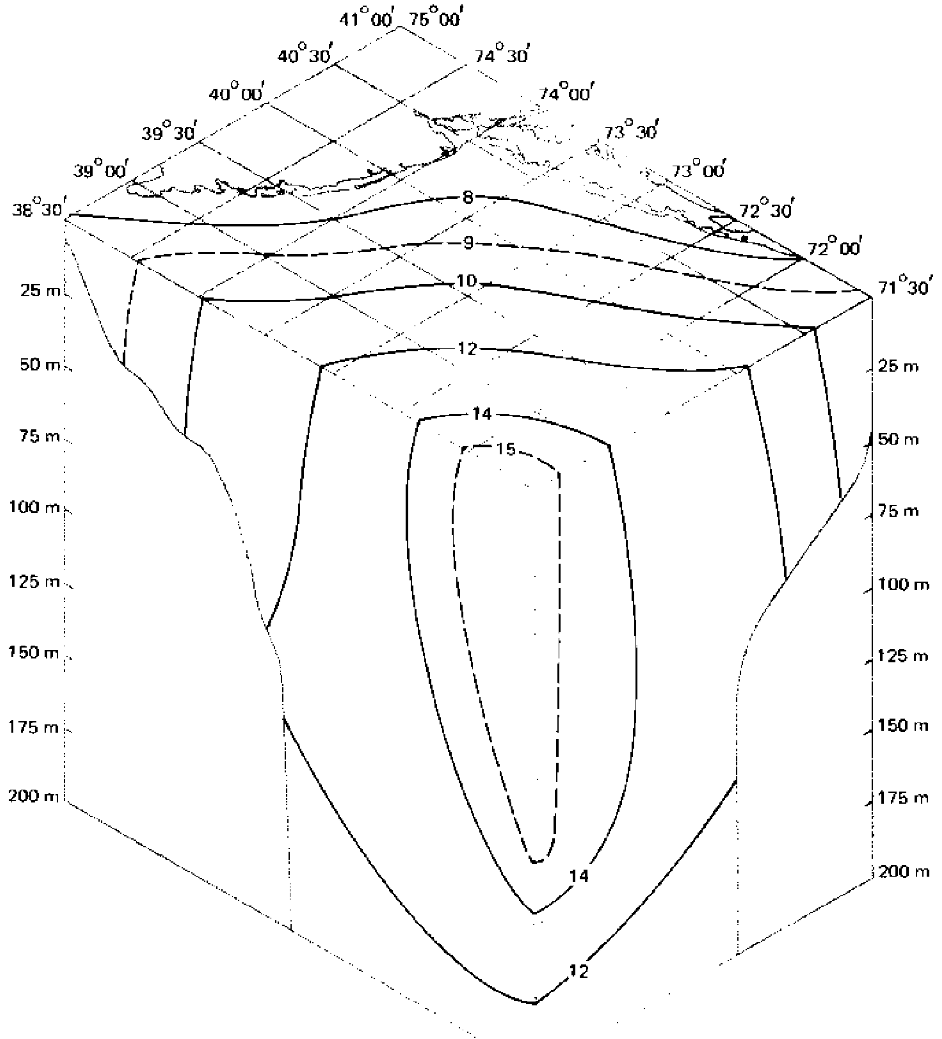
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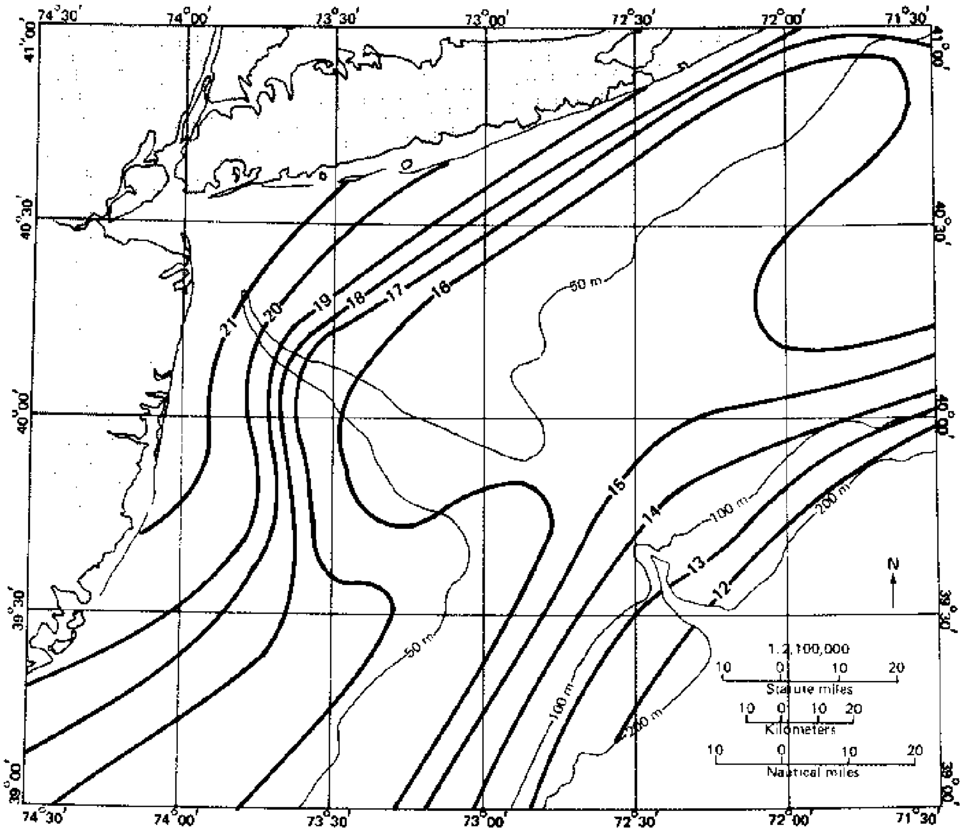
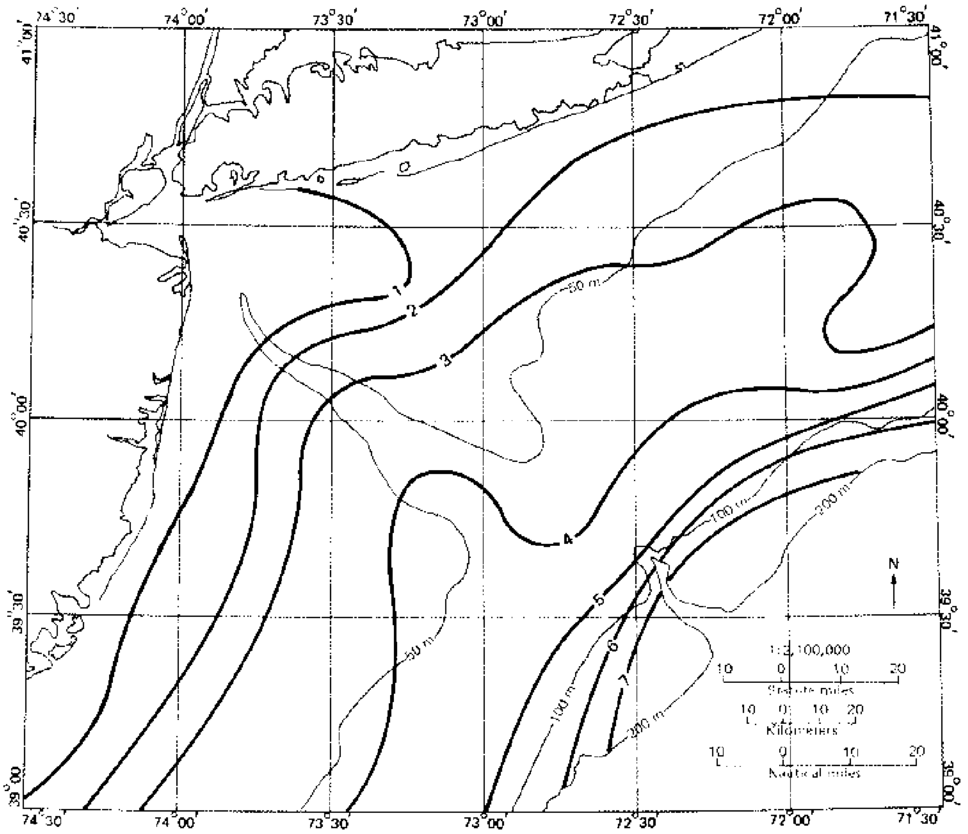
Map 5. Mean extreme bottom water temperature

Maximum



Minimum



Units are °C

Source: After Colton and Stoddard 1973

Salinity Distribution

The seasonal cycle of salinity in the Bight is not nearly so well defined as the temperature cycle and often has an amplitude less than the year-to-year variability. The observed salinity distribution on the continental shelf is a consequence of the balance between the influx of river runoff, evaporation minus precipitation, and the advection and mixing of saltier slope water from offshore. Whether there is an annual cycle in the influx of slope water onto the shelf is not well documented, although Wright (1976) suggested that exchange is greatest during late summer and early autumn.

Salinity reaches its seasonal maximum at the end of winter because of subfreezing conditions on the continent and drops to its minimum by early summer because of the spring river runoff peaking in April. Throughout autumn, vertical stirring is strong, and mixing of slope water onto the shelf more than counterbalances dilution from river runoff, slowly increasing salinity from the summer minimum toward the winter maximum.

The salinity of the Bight is described here in three ways. First, the data averages for all years and all months for each 1° Marsden square were calculated.

Second, the seasonal cycle is illustrated with salinity diagrams from individual cruises. The apex maps were produced from data gathered during numerous cruises by Middle Atlantic Coastal Fisheries Center (1972) personnel in 1969 and by MESA staff (Hazelworth et al 1974; Hazelworth and Weiselberg 1975) in 1973-74. The vertical sections, selected from Bigelow and Sears (1935), are oriented along a line from the Hudson estuary mouth southeast across the shelf in the vicinity of Hudson Shelf Valley (see locator map in Figure 6). Because gradients in the salinity structure are vertical and cross-shelf, these sections illustrate most of the important characteristics of seasonal salinity on the shelf.

Third, salinity variations are useful tracers for depicting water movements. Existing data are sufficient to provide a limited understanding of the mixing of riverine and nearshore coastal waters as well as shelf/slope exchange processes.

Mean Salinity

As seen in Map 6, mean contours of salinity on the shelf follow the bathymetry and show little along-shelf variability throughout the Bight. Salinity steadily increases seaward from less than 32°/oo along the coast to between 32 and 34°/oo over the central shelf, reaching approximately 35°/oo at the shelf break. As a result of the freshening of coastal waters from river runoff, an offshore component to the prevailing along-shelf residual surface drift exists (Bumpus 1973). The associated coastward intrusion of more saline bottom waters causes the isohaline surfaces to intersect the shelf bottom obliquely. Consequently, vertical gradients are typically about 0.025 to 0.05°/oo m^{-1}. Evidence of the subsurface salinity maximum at 100 m can be seen on the southeastern vertical edge of the block diagram in Map 6.

The salinity inversion and associated surface front persist throughout the year (Bigelow and Sears 1935; Bowman and Weyl 1972; Boicourt 1973; Beardsley and Butman 1974; Flagg and Beardsley 1975; Beardsley et al in press; Boicourt and Hacker 1976; Gordon et al in press). In winter the salinity and temperature inversions coincide. A strong seasonal thermocline dominates surface waters in summer and replaces the temperature inversion, but the salinity intrusion slopes upward through the thermocline to the surface over the continental slope.

Seasonal Cycle (Map 7, Figure 6)

Winter (January, February, March). Salinities in the Bight peak in late February or early March when surface values reach 33 to 34°/oo on the inner shelf. There is a small increase seaward, with values of about 34°/oo at the shelf edge. The vertical salinity gradient is smallest in winter—typically 0.005°/oo m^{-1}; normally the highest salinities are at the bottom, usually about 33°/oo near shore, increasing to about 35°/oo at the shelf break. In the February 1929 vertical section in Figure 6, the sharp front centered around 34.5°/oo delineates the top surface of the salinity maximum intersecting the shelf around 100 m.

In the Bight apex, surface salinities in January/February rise to their annual maxima of greater than 34°/oo (Map 7). Bottom salinities follow the bathymetry at all depths and are greater than 34°/oo over most of the apex. Salinities begin to drop in March as the spring thaw commences; typical values are 29 to 32°/oo.

Spring (April, May). In April, dilution from river runoff is most pronounced closest to shore; salinities away from the direct influence of river mouths often drop to below 33°/oo near shore. During May a low-salinity band less than 32°/oo develops along the whole Bight coastline. This 32°/oo isohaline meanders seaward near the mouth of Long Island Sound and the Hudson River where it is occasionally pushed up to 150 km offshore.

The effects of river runoff and penetration of slope water tend to increase the vertical salinity gradients, which can reach as high as 0.4°/oo m^{-1} near estuary mouths and usually are steepest in the top 20 m. However, the observed characteristics vary from location to location and from year to year. In Figure 6 the May 1930 and May 1931 vertical profiles illustrate how distributions can differ from one year to the next, indeed over a few days. The May 1930 profile shows a rather steady increase in salinity seaward, whereas a year later (17-18 May 1931) a very strong salinity front with contrasts of almost 2°/oo is present over the shelf edge.

Salinity in the apex is strongly influenced by the Hudson River spring runoff, which produces large horizontal and vertical gradients in salinity and density. The position of the well-developed river plume is apparently sensitive to wind stress and reversals in residual drift on the shelf because its axis can swing from south along the New Jersey shoreline to east along the south shore of Long Island (Map 7); at other times the plume divides into southerly and easterly components (Redfield and Walford 1951). Bottom salinities generally follow the bottom contours. A strong salinity gradient appears at the mouth of the Hudson estuary as river runoff peaks in April.

Vernal freshening off the mouth of the Delaware River results in a localized lowering of salinity that peaks in May. During spring, indrafts of slope water sometimes penetate shelf waters, usually near the bottom. This phenomenon varies from year to year, but in extreme cases the effects of vernal freshening near shore may be entirely counterbalanced by these inflows of slope water (Bigelow and Sears 1935).

Summer (June, July, August). The salinity band less than 32°/oo remains almost stationary during June, July, and August. Its outer edge is usually located over the central shelf (Figure 6), although occasionally it moves seaward. High vertical gradients are found over most of the shelf, with large seasonal values over the central and outer shelf, especially at the salinity front.

Surface salinities in the apex range from about 27°/oo near the Sandy Hook-Rockaway transect to about 30 to 31°/oo at the southeast corner (Map 7). The Hudson plume contracts and often breaks into isolated pools.

Bottom salinities in the apex range from 27 to 29°/oo along the Sandy Hook-Rockaway transect to 30 to 32°/oo at the outer edge.

Autumn (September, October) and Early Winter (November, December). Destabilization by heat and momentum exchange during autumn is usually stronger than buoyancy from river runoff. This stirring reduces vertical salinity gradients at a rate that changes considerably from year to year and leads to a steady increase in surface salinities—often as large as 0.8°/oo from July to October.

In the October 1931 vertical section in Figure 6, the effects of decreased stability on the shelf are indicated by the steepening isohalines. The slope front distinctly separates shelf and slope waters, and the underlying salinity maximum intersects the shelf around 130 m.

Surface salinities continue to increase through November and December until the winter maximum is attained in January. Strong winter vertical stirring dissipates any isolated patches of low-salinity waters formed during the summer. Subsurface salinities also increase during autumn to 33°/oo in Hudson Shelf Valley (Map 7). The implication is that advection of slope water onto the shelf must exist to counteract the inflow of river water. Both are mixed vertically, resulting in gradients less than 0.005°/oo m^{-1} at depth over most of the shelf during December.

In summary, apex salinity closely follows the annual cycle of runoff and surface salinity at Ambrose Light Tower (Chase 1969). Near the southeastern corner of the apex a two-month delay is evident; salinities peak in January and drop to a seasonal minimum in June. This is comparable to coastal conditions away from the direct influence of river outflow (Ketchum and Keen 1955; Howe 1962; Ketchum and Corwin 1964).

Cross-Shelf Exchange Processes

The following discussion on salinity variability refers to the standard deviation of processed NODC data.

From time to time extremes in salinity will be found that may be much greater than the standard deviations indicate. However, for a normal distribution, 68% of a large number of observations spread over many years and seasons would be expected to lie within ±1 standard deviation and 95% of the observations within ±2 standard deviations.

Map 8 shows that surface variability is typically less than ±2°/₀₀ along the coast, on the order of ±1°/₀₀ over most of the shelf, and decreases to ±0.75°/₀₀ at the southeastern corner of the Bight. Thus, the greatest salinity variability is near shore, reflecting seasonal land runoff patterns.

In the inner Bight, variability in the Hudson River runoff produces bottom values greater than ±1°/₀₀. Slope intrusions into bottom shelf waters are responsible for the increase in variability seaward from the central shelf low-variability band (±0.50°/₀₀ at 40 to 50 m); a maximum of about ±0.85°/₀₀ occurs near 100 m. Decreasing variability over the slope indicates the stability of deep Atlantic water.

Ketchum and Keen (1955) studied the concentration of fresh water on the continental shelf and its relation to river runoff between Cape Cod and Chesapeake Bay. Assuming no northern input of water via Nantucket Shoals, they concluded that the total volume of fresh water in winter and spring was equal to that discharged by the rivers in 18 months.

Away from river mouths, there is a two- to three-month delay between runoff and the corresponding adjustment in shelf salinity (Howe 1962; Ketchum and Corwin 1964). This mixing time allows an extra 25% of fresh water to accumulate in summer, because two successive spring runoffs are present on the shelf at the same time.

Ketchum and Keen also found a decrease in the mean content of fresh water southward in spite of the addition of water along the coast. They concluded that cross-shelf exchange of water and salt must necessarily increase southwards. They calculated that, regardless of depth, salt is transported offshore (positive) between spring and summer and onshore (negative) between summer and winter (Table 2).

From a study of cross-frontal mixing, Voorhis, Webb, and Millard (1976) calculated that approximately 5×10^{15} g of salt are annually added to the Middle Atlantic Bight from slope water.

This input of salt is sufficient to form new shelf water, of salinity 33°/₀₀, when mixed into the annual mean river influx of 150 km^3 yr^{-1} (Bué 1970).

Wright and Parker (1976), in a temperature-salinity volumetric census of the Middle Atlantic Bight, found little change in the total volume of shelf water (water with salinity less than 35°/₀₀) between summer and winter. Almost half the shelf water in the Bight is located in surface layers beyond the shelf break and above the slope front. As Ketchum and Keen had assumed, Wright and Parker found little evidence that unmodified Gulf of Maine water reaches the Bight via Nantucket Shoals.

Boicourt (1973) and Boicourt and Hacker (1976), who extensively investigated slope water intrusions off Chesapeake Bay, found that strong northward winds can drive an offshore *Ekman surface flux*, accompanied by a return flow of warmer, saltier slope water, onto the shelf during both summer and winter. In winter the return flow transports slope water onto the shelf along the bottom, stratifying the mid-shelf region and resulting in weak stratification

Table 2. Mean salinities and flux of salt onto continental shelf between Cape Cod and Chesapeake Bay

Depth Contour	Mean Salinities (°/₀₀)			Flux of Salt (10^7 g/sec)	
	Spring	Summer	Winter	Spring-Summer (offshore)	Summer-Winter (onshore)
m					
37	32.54	32.24	32.64	2.53	2.24
55	32.88	32.40	32.84	7.86	5.48
73	33.24	32.69	33.18	14.68	9.52
183	34.06	33.57	34.04	25.78	16.58
1,830	34.88	34.86	34.91	30.03	23.63

NOTE: Contours average 28 km apart

Source: From Ketchum and Keen 1955

throughout the water column. Unfortunately, inadequate data exist to determine whether there is a *relaxation* or whether the high-salinity intrusion is locally stirred and mixed when winds return to their prevailing winter southward direction.

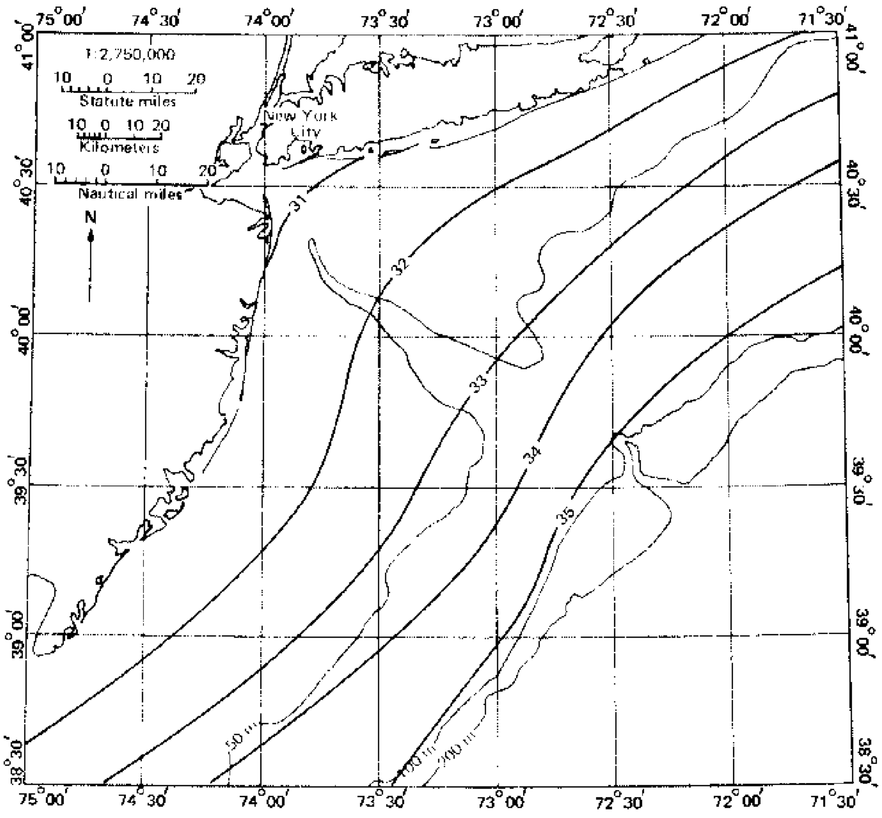
Prevailing winds in summer are northward, opposite to the mean flow (about 5 cm sec^{-1}) of shelf water (Bumpus 1969). These winds also give rise to an offshore Ekman flow in the surface waters, with, according to Boicourt and Hacker, a return flow of slope water at mid-depth (20 m) in the thermocline, not as a bottom Ekman layer as in winter. These intrusions, often only a few meters in vertical extent, have been observed to penetrate shelf waters above the cold pool during August.

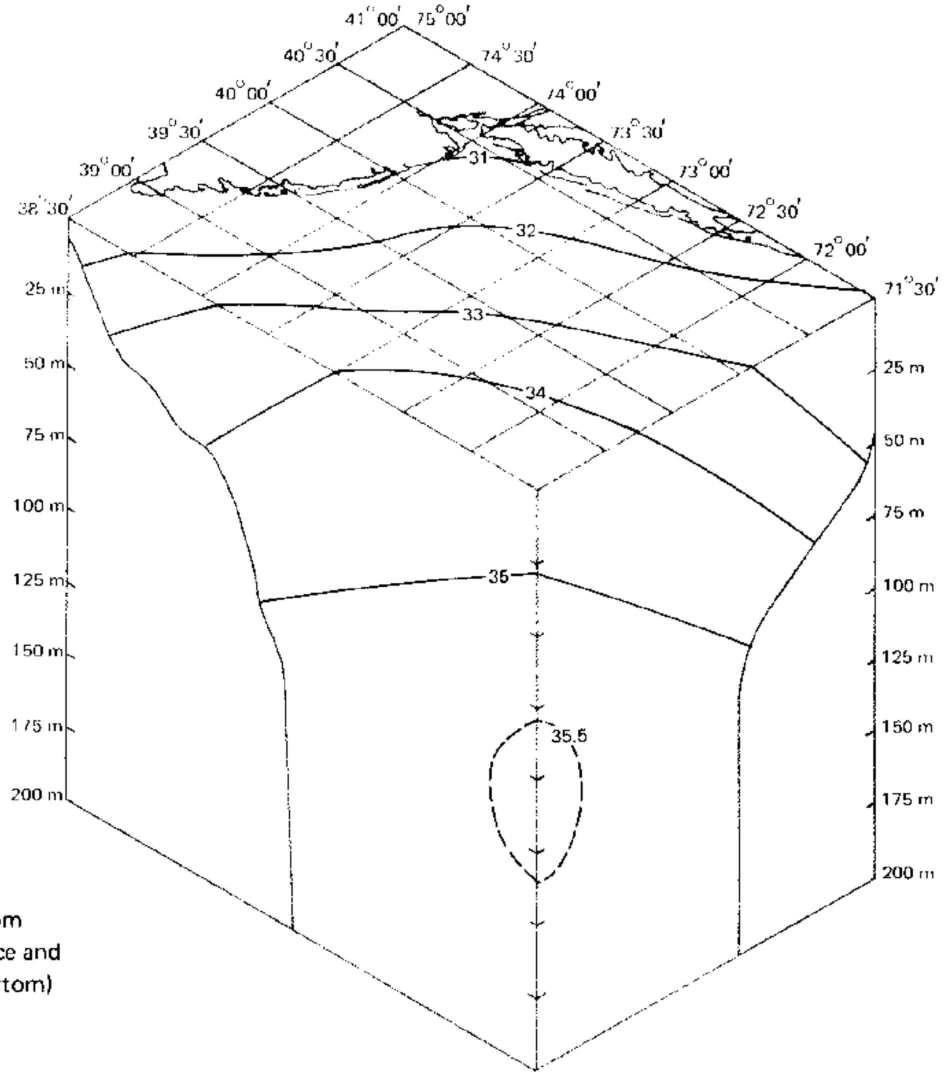
Gordon et al (in press) also found evidence of slope water intrusions along isopycnal surfaces in the *pycnoline* of New York Bight shelf during October 1974, suggesting that such intrusions may be widespread in the Middle Atlantic Bight. Breaking internal waves generated at the shelf edge by semidiurnal and diurnal tides (Apel et al 1975) may also stir and mix slope water into outer shelf waters.

Beardsley et al (in press) found surprising consistency among three estimates (approximately 8 x 10^{12} m^3 yr^{-1}) of along-shelf coastal transport taken at cross sections off New England, New York, and Chesapeake Bay. This, with an estimated shelf water volume of 6 x 10^{12} m^3, leads to a residence time of 0.75 yr, somewhat less than Ketchum and Keen's (1955) 1.5 yr (they assumed no replenishment from the north).

In summary, the relative importance of salt and water fluxes into the Middle Atlantic Bight from the Gulf of Maine and from cross-shelf exchange with slope water remains unclear and elusive. Studies of cross-shelf transport must take into account the prevailing southwesterly and southerly residual drift throughout the Bight (Bumpus 1973), reversals in the coastal current (Bumpus 1965), and high-velocity flows (60 cm sec^{-1}) driven by strong winds (Beardsley and Butman 1974). Strong southward winds are capable of transporting outer shelf water one shelf width in only three days (Boicourt and Hacker 1976). Winds over the Bight are highly time dependent—variability exceeds the mean 5 to 10 times (Stommel and Leetmaa 1972); intrusions are intermittent, and intensities vary greatly even over a few days. Much more intensive synoptic sampling will be necessary to quantify shelf/slope exchange and its seasonal variability.

Map 6. Annual mean salinity distribution

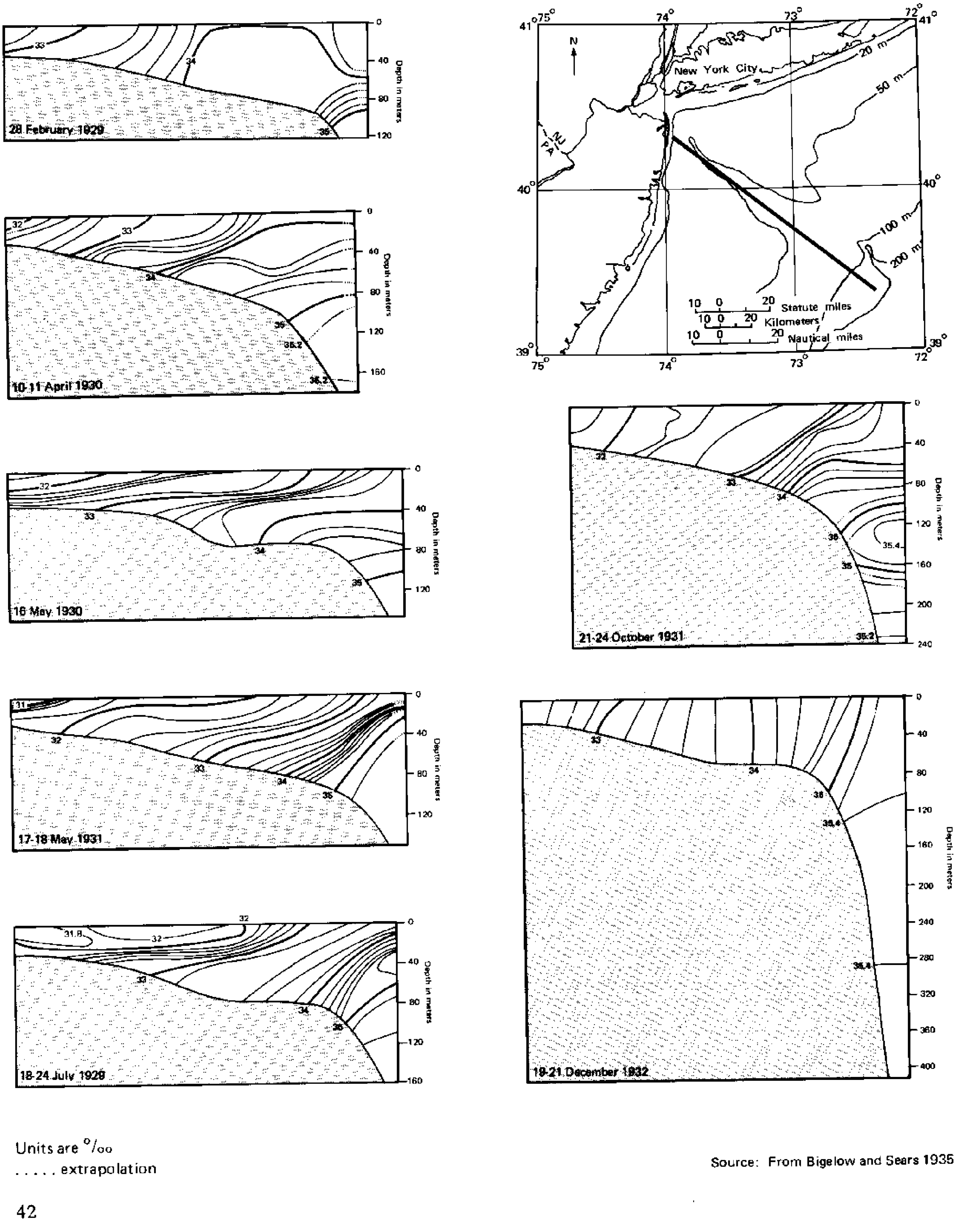




Map pairs show bottom contours (top), surface and vertical contours (bottom)

Units are °/oo

Figure 6. Salinity profiles across continental shelf



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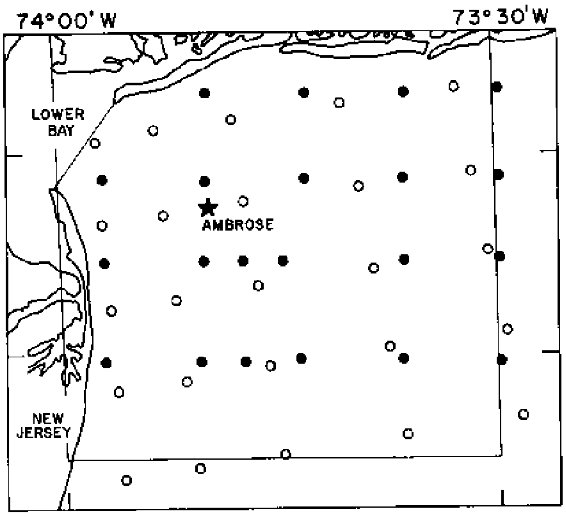
Source: From Bigelow and Sears 1935

Map 7. Salinity distribution, Bight apex, 1969 and 1973



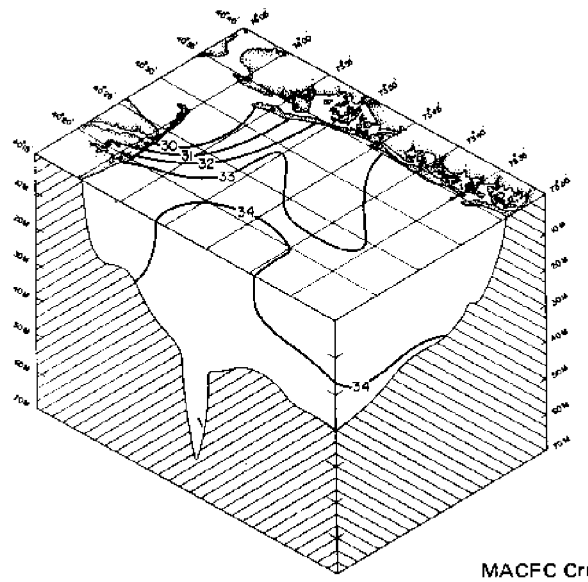
Map to left shows station locations
- ○ MESA 1973-1974
- ● Middle Atlantic Coastal Fisheries Center (MACFC) 1969

Map pairs show surface and vertical contours in block diagram, bottom contours in map

Units are ‰

Data Sources: Middle Atlantic Coastal Fisheries Center 1972; Hazelworth et al 1974; Hazelworth and Weiselberg 1975

6-7 FEBRUARY 1969



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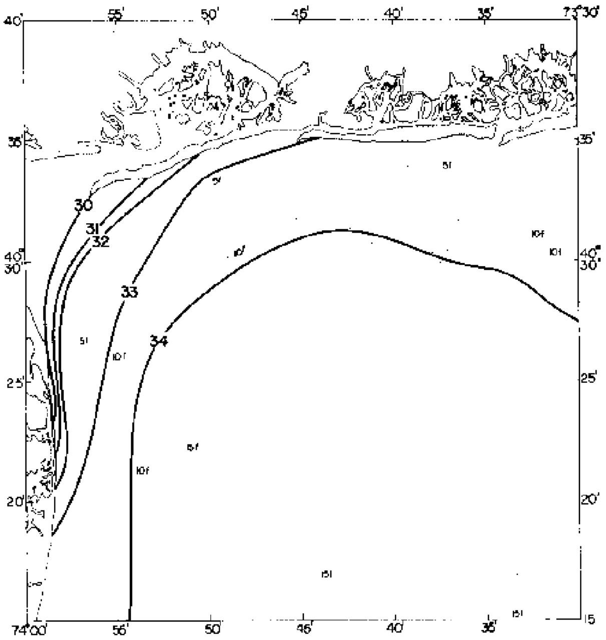

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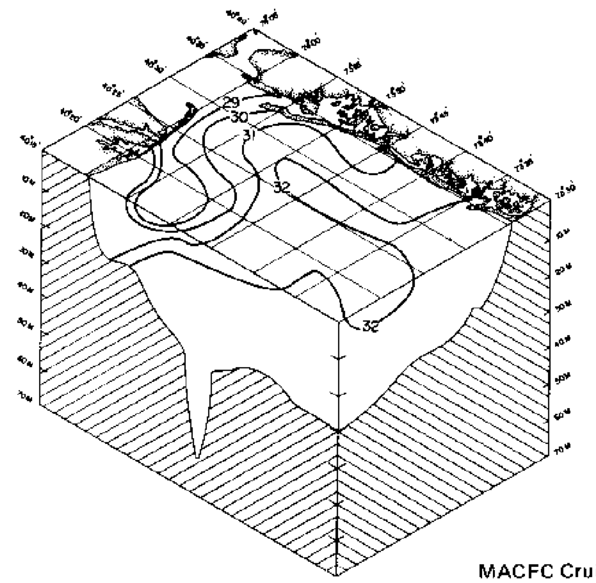

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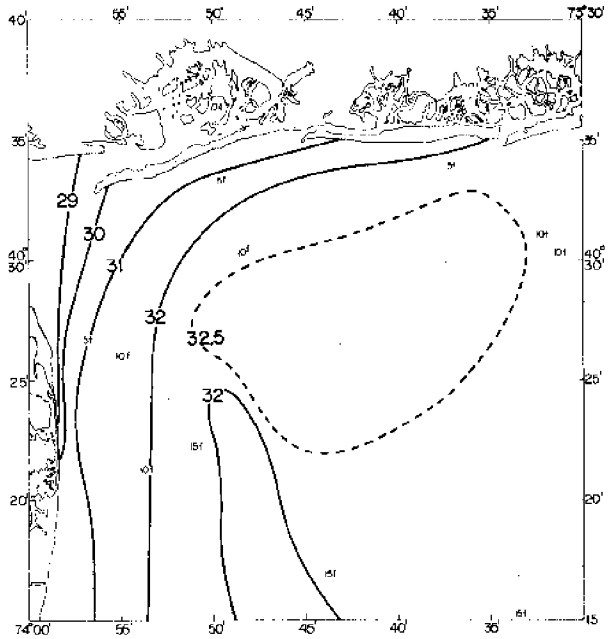

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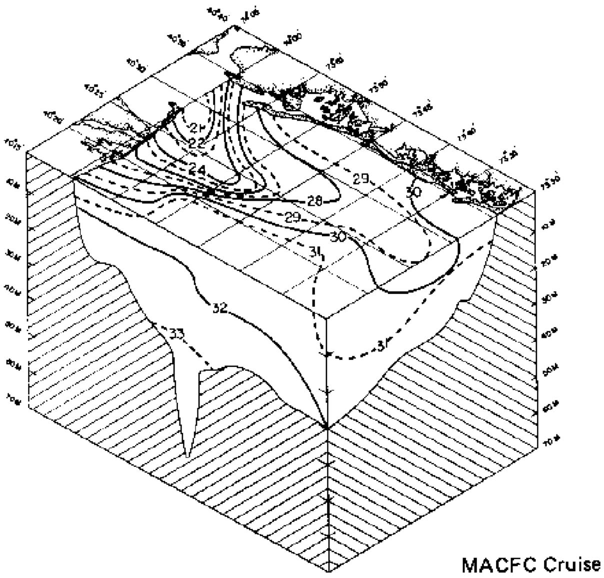

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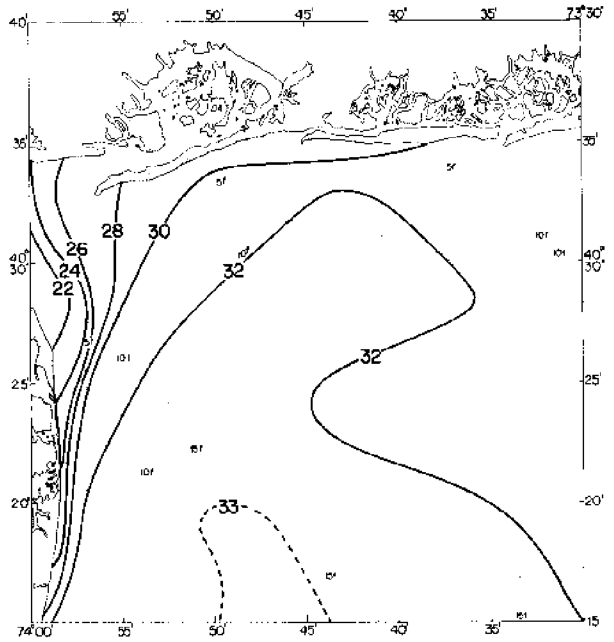

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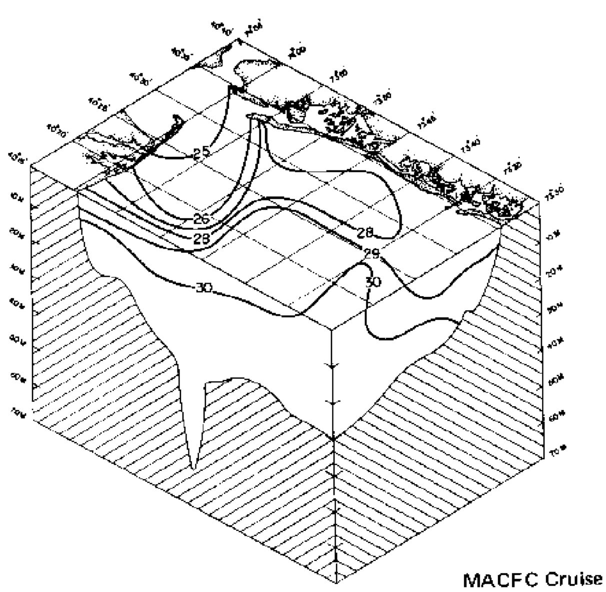

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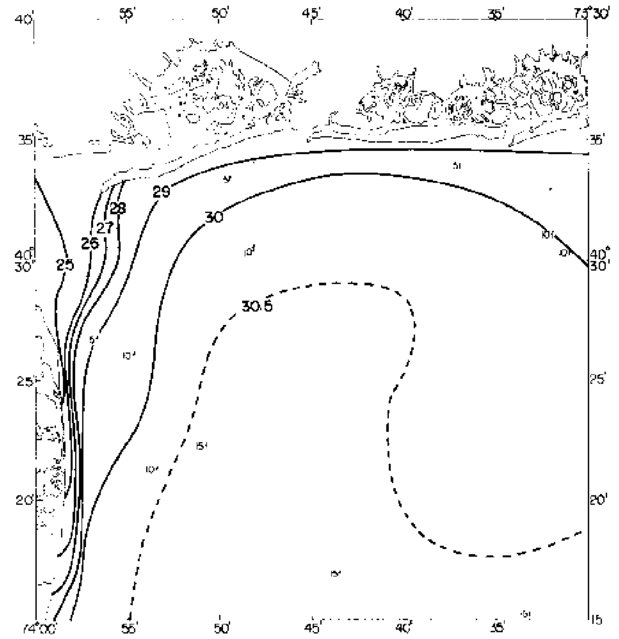

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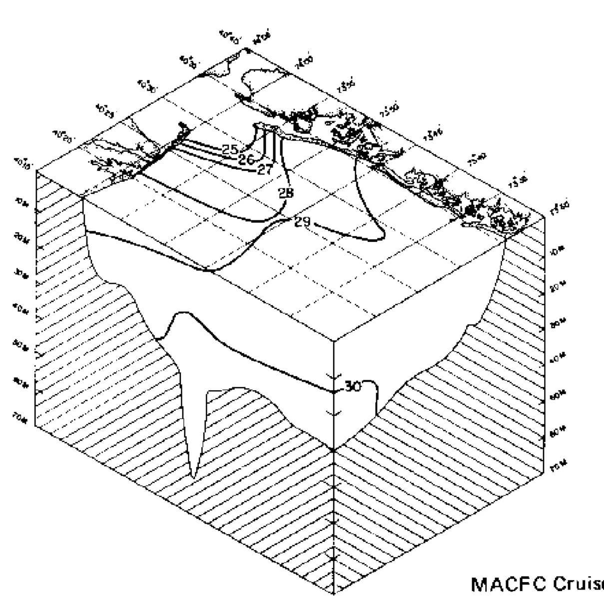

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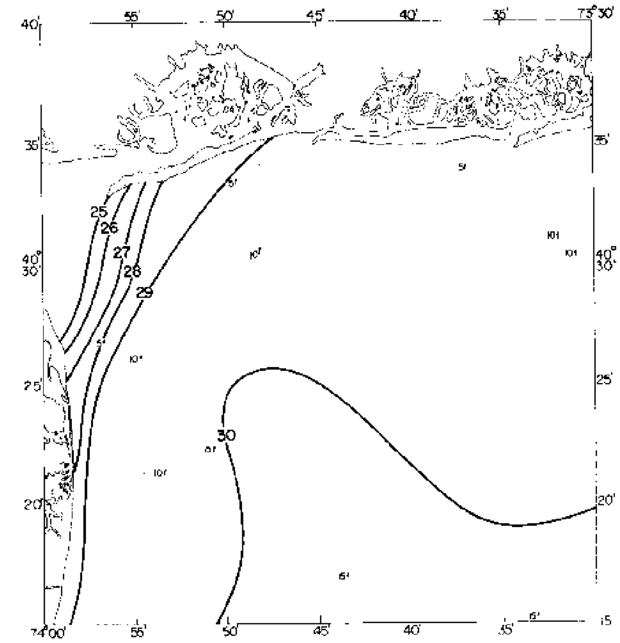

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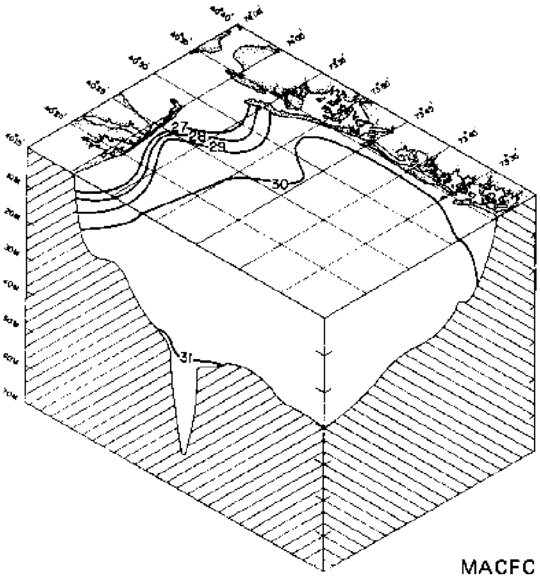

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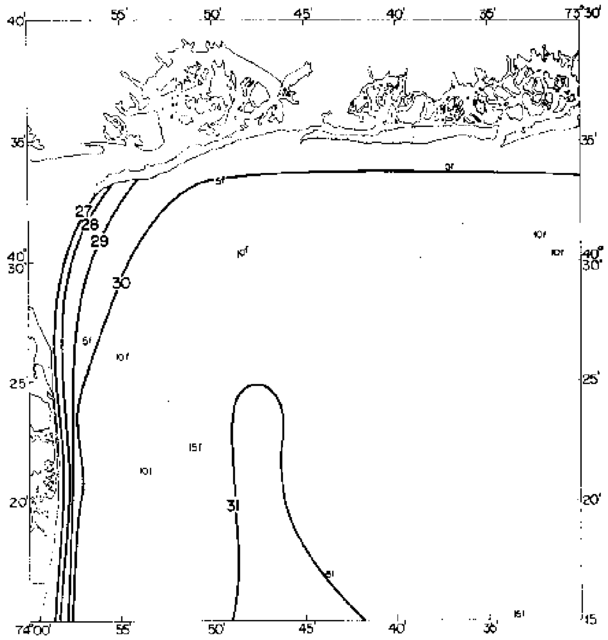
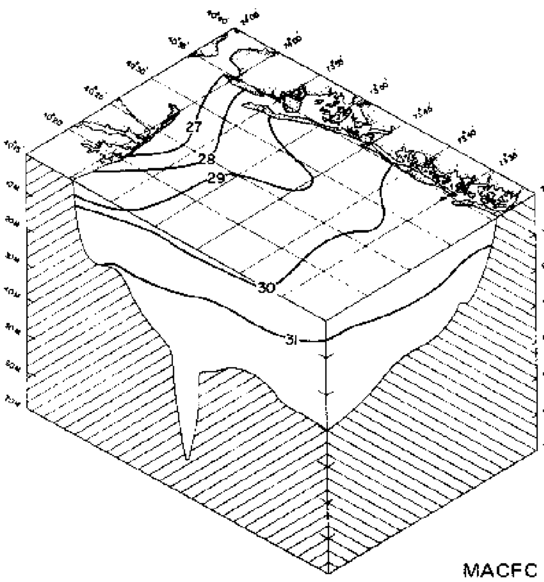


28 JULY-1 AUGUST 1969



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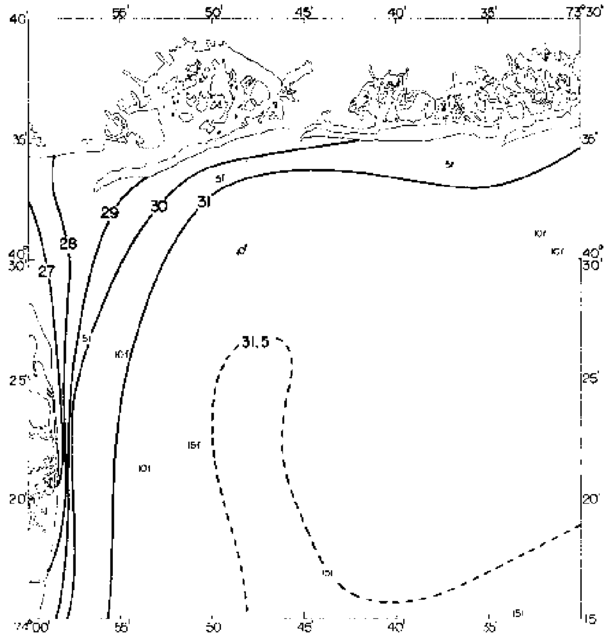

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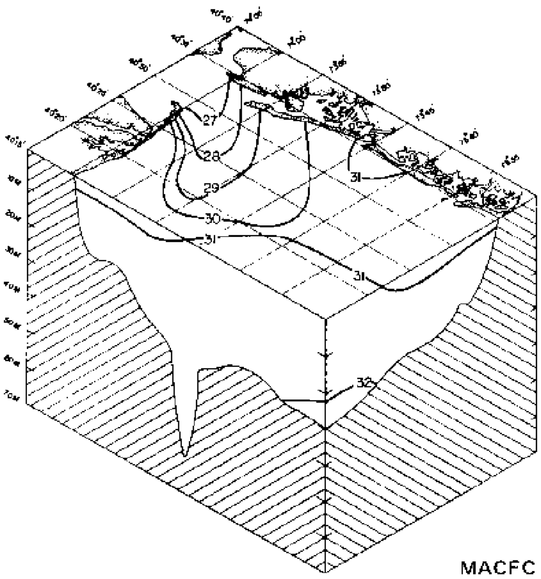

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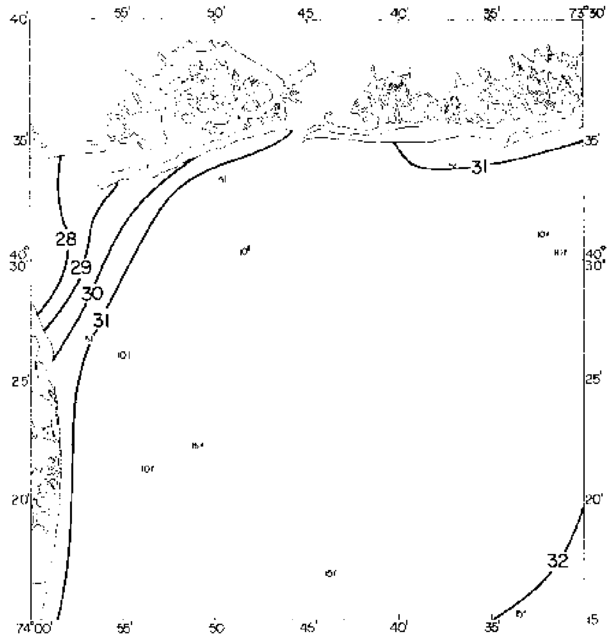

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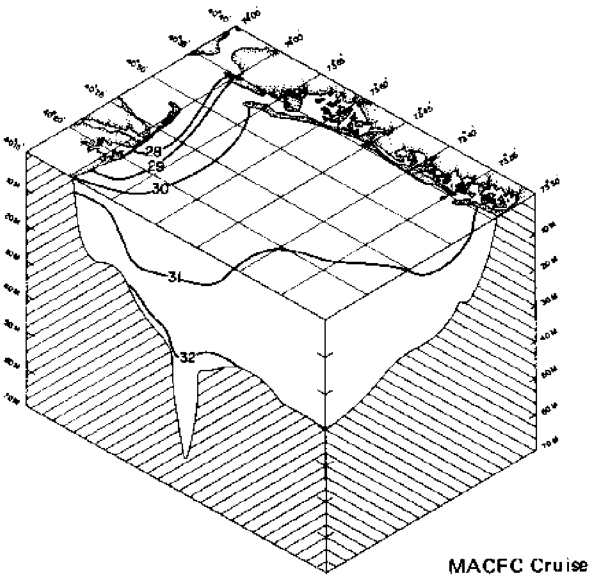
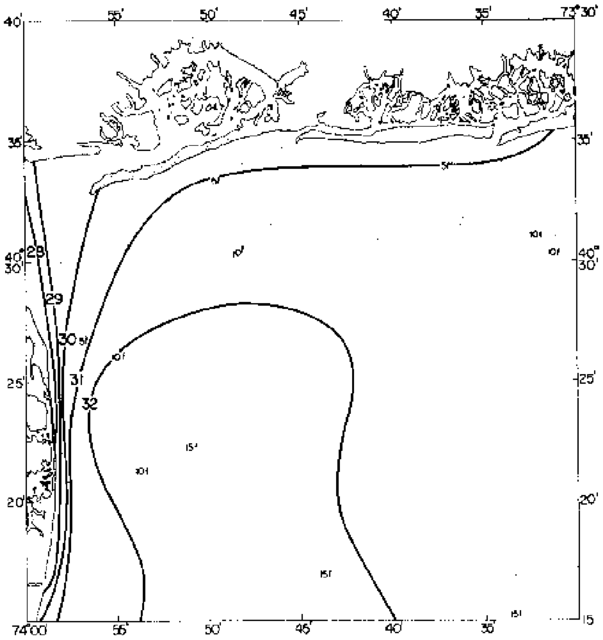
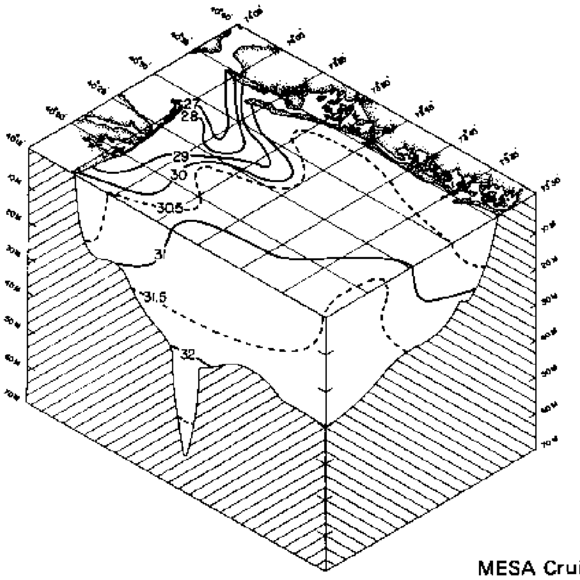


MACFC Cruise 10



1-4 OCTOBER 1973



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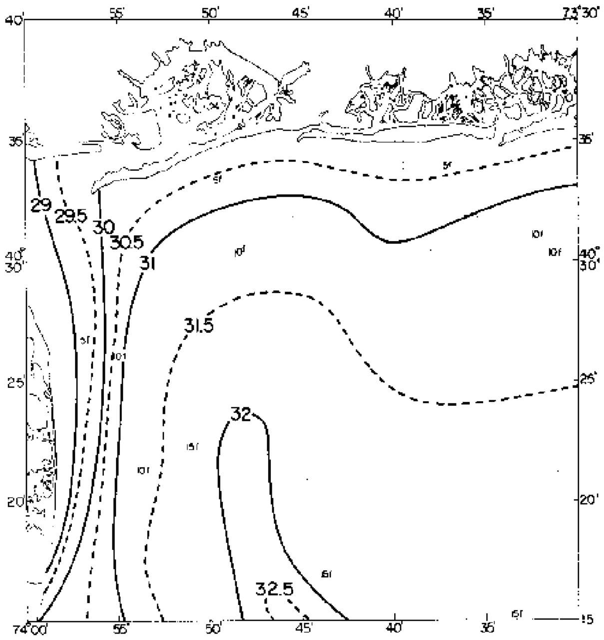
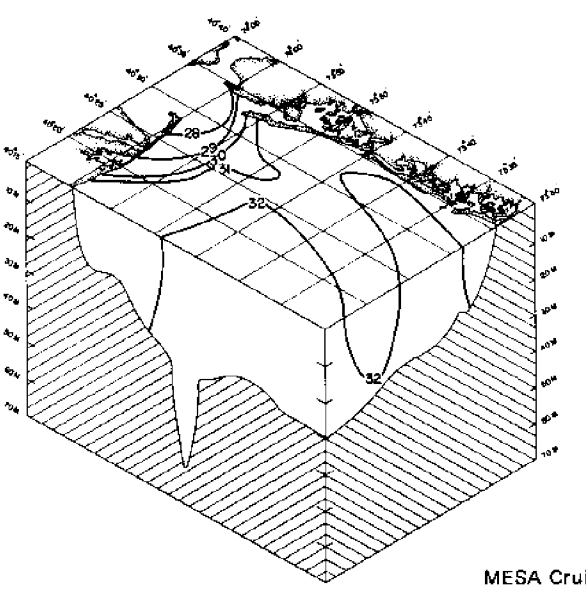


5-9 NOVEMBER 1973



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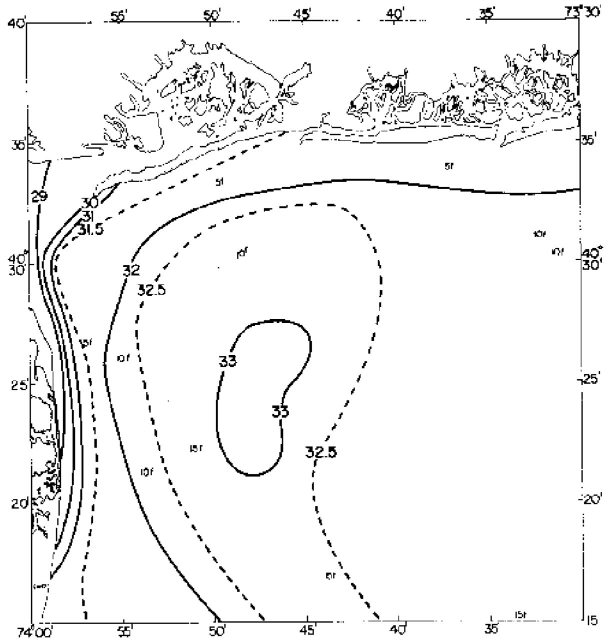
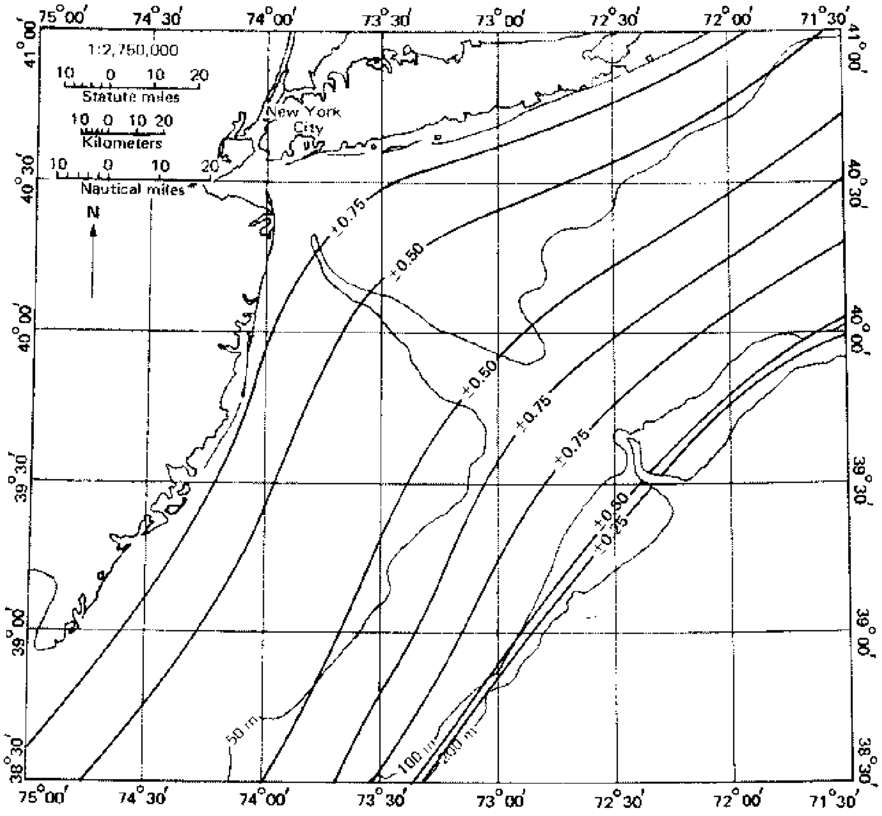
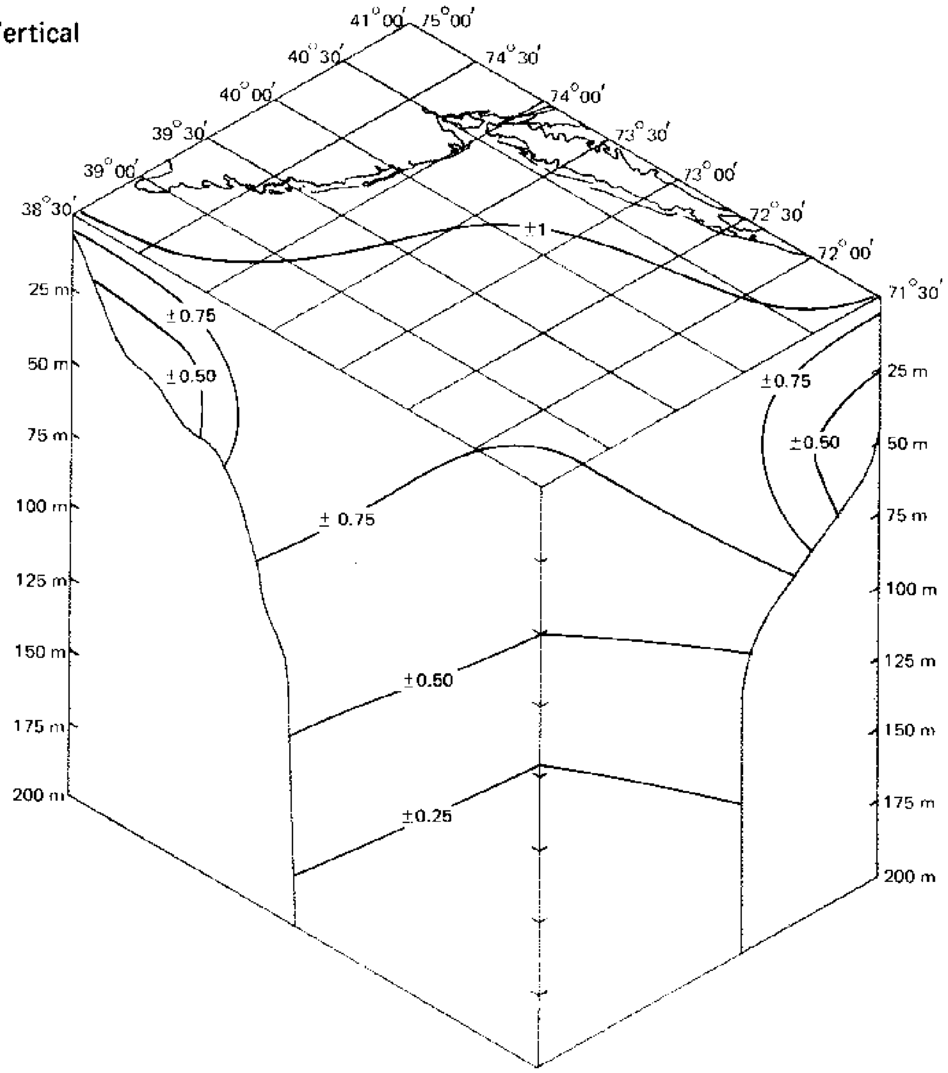


Map 8. Salinity variability

Bottom



Surface and Vertical



Units are °/oo

Density Distribution

The distribution of density anomalies in the Bight follows a definitive seasonal cycle. Isometric and bottom contours of sigma-T, representing averages over five seasonal intervals, are used since monthly progressions were often obscured by deviations from year to year, principally affected by corresponding deviations in salinity. Also, data were lacking for some winter months, particularly January.

Winter Maximum
(Maps 9 and 10—January, February, March)

Bight water reaches its maximum density during winter for two reasons: temperatures descend to their yearly minimum and salinities attain their annual maximum. Sigma-T contours for January-March are almost vertical over the shelf (Map 10). Surface values are about 24 in the Bight apex and increase steadily to the shelf break. Horizontal gradients then diminish over the slope and there is a gradual vertical increase in density from 26.5 at the surface to 27 at 200 m. Vertical isopycnals on the shelf are replaced by horizontal contours over the slope as a consequence of permanent stratification in the deep ocean. Often the contributions to the density within the temperature and salinity inversions under the slope front cancel each other out (Bowman and Weyl 1972; Boicourt 1973; Gordon et al, in press) and no sharp density gradients separate shelf and slope water. However, during late winter, Flagg and Beardsley (1975) and Beardsley and Flagg (1976) found a sharp density front over the New England shelf break, dominated by a corresponding salinity front that more than counterbalanced the coincident temperature front.

Spring
(Maps 9 and 10—April, May)

During vernal warming, the large increase in river and groundwater runoff and the simultaneous warming of surface waters produce a marked stratification of the water column. This leads to a density distribution of sloping isopycnals over the shelf and increases the steric anomaly (Meade and Emery 1971) that drives both the southward coastal current (Bumpus 1965; Bumpus and Lauzier 1965; Stommel and Leetmaa 1972) and the two-layered cross-shelf transport (Bumpus 1973).

The sigma-T surfaces continue to remain almost horizontal over the southeastern corner of the Bight slope waters. Surface values range from 17 to 24 in the apex to 25.8 over the slope, and then increase steadily to 27 between 200 and 1,000 m (Stommel 1965). Surface and bottom contours generally follow the bathymetry at all depths; vertical gradients are 0.02 sigma-T units m^{-1} over most of the shelf.

Summer Minimum
(Maps 9 and 10—June, July, August)

The transition from vertical to horizontal density stratification is complete by the end of June, and the density field is dominated by the strong summer thermocline except near the New Jersey coast where the Hudson plume is influential. Surface densities throughout the Bight drop to their annual minima—from 20 in the apex to over 23 at the Bight's seaward boundary. Most of the vertical stratification on the shelf lies in the top 50 m of the water column where vertical gradients are maximum for the year—typically 0.06 to 0.10 m^{-1}. Horizontal gradients at the bottom are greatest on the inner shelf where there is a rapid increase from 18 near the Sandy Hook-Rockaway transect to 26 at about 50 m. These sharp gradients are the consequence of the intersection of the summer thermocline with the bottom. Beyond the 50 m contour bottom densities increase slowly to the shelf break where sigma-T is 27. Isopycnal spreading and mixing occurs in this zone during summer (Boicourt 1973; Boicourt and Hacker 1976; Gordon et al, in press).

Autumn
(Maps 9 and 10—September, October)

The erosion of the summer thermocline is reflected in the September-October density profiles in Map 10 with the descent of the sigma-T = 26 isopycnal from its 50 m summer depth to about 75 m. Surface densities have increased slightly due to both a decline in temperature and an increase in salinity; the combined effect is a reduction in vertical gradients in the surface layers to about 0.03 to 0.04 m^{-1}. Bottom densities continue to increase steadily seaward from about 20 to 24 in the apex to 24 to 26 over the central shelf. The 26 contour has migrated seaward about 100 km because of a density reduction from the increase in bottom temperatures from around 7°C to about 12°C in October.

Early Winter
(Maps 9 and 10—November, December)

In winter, a homogeneous water column replaces the summer stratification as vertical equalization continues. Surface densities continue to increase as temperatures decline and salinities increase toward their late January extrema. Values range from 24 in the apex to about 25.7 at the outer edge of the Bight. Contours follow the coastline at all depths; bottom contours range from 24 in the apex to 27 at the 200 m isobath.

Temperature-Salinity Relationships

Winter. A 1974 winter temperature-salinity (T-S) diagram for the area between the Hudson Canyon and Nantucket Shoals is shown in Figure 7 (Flagg and Beardsley 1975). Four distinct regions are evident. In region 1 the curves split into two components. The lower fork is formed of low-salinity, cold, inshore water; the upper fork is due to the presence of anomalously warm water from Long Island Sound.

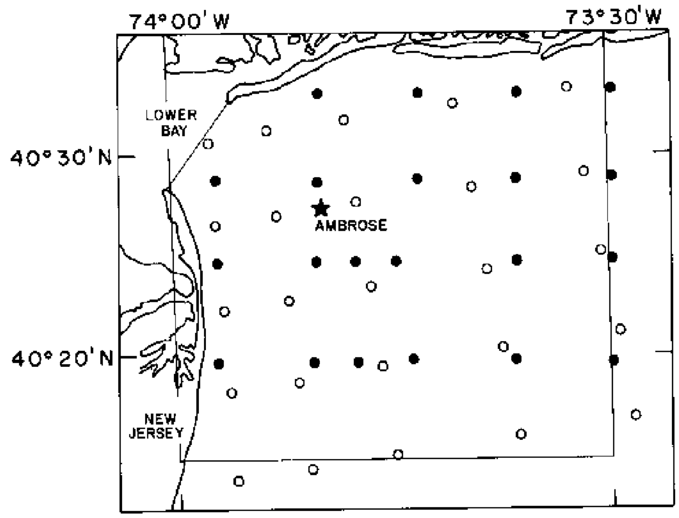
Region 2, the principal segment (5.5°C, 32.8°/oo to 13.5°C, 35.5°/oo) of inner and central shelf water, including the frontal zone, indicates relatively simple and uniform mixing of nearshore and offshore water.

Region 3 (13.5°C, 35°/oo to 12°C, 35.5°/oo) is composed of surface slope water outside the front and down to 200 m. Region 4 (12°C, 35.5°/oo to 4.5°C, 34.98°/oo), composed of slope water below 200 m, has no seasonal cycle and is typical of North American Basin water (Wright and Worthington 1970). In its simplest interpretation this T-S correlation indicates three-point mixing of inner shelf water, surface slope water, and North Atlantic central water.

Summer. Mixing of shelf and slope water in summer is illustrated with data from two transects (Figure 8) across the Bight during August 1971 (Bowman and Weyl 1972). Surface waters are almost isothermal around 24°C. Bottom shelf waters are isolated from surface waters by a strong thermocline; horizontal mixing spreads along isopycnal surfaces at sigma-T = 25.5. A vertical profile taken over the slope at station 95 (Figure 8) includes an intrusion of 35°/oo water at 25 m and the temperature/salinity maximum at 125 m. North American Basin water (Wright and Worthington 1970) is shown for comparison.

In an October 1974 study by Gordon et al (in press) a careful T-S analysis of Bight slope water determined the presence of North Atlantic central (deep slope) water, Irminger Atlantic water (North Atlantic central water originating south of Iceland, 300 to 1,000 m), Labrador Sea water (1,000 m), a Mediterranean component of North Atlantic deep water (2,000 m), and North Atlantic deep water (over 2,000 m).

Map 9. Density distribution, Bight apex, 1969 and 1973



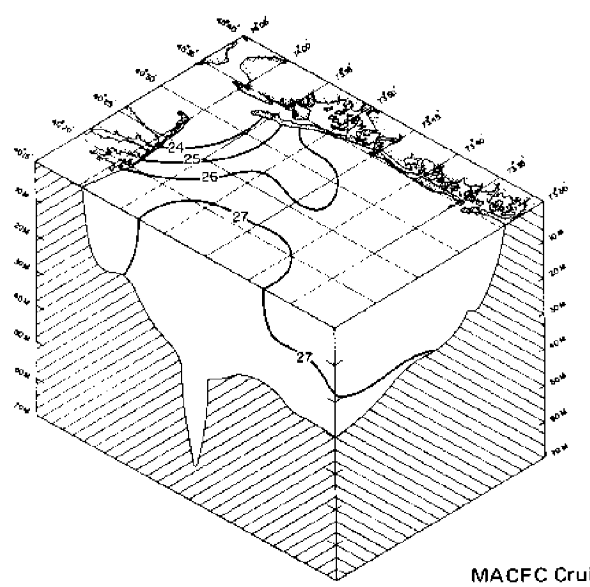
Map to left shows station locations
○ MESA 1973-1974
● Middle Atlantic Coastal Fisheries Center (MACFC) 1969

Map pairs show surface and vertical contours in block diagram, bottom contours in map

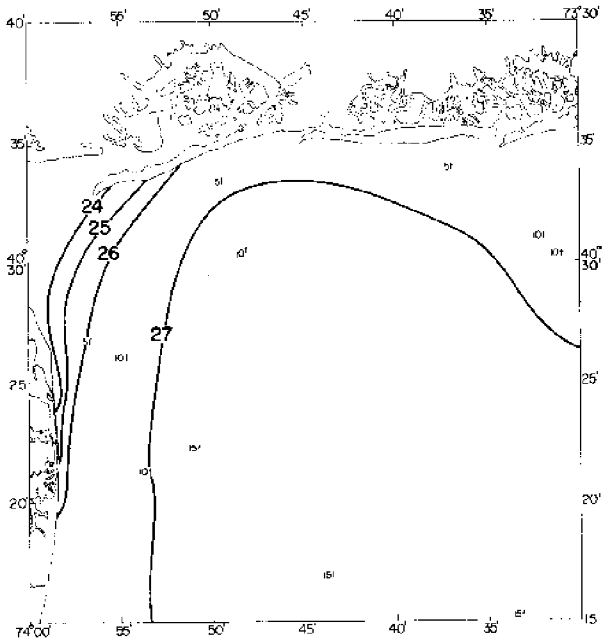
Units are σ_T

Data Sources: Middle Atlantic Coastal Fisheries Center 1972; Hazelworth et al 1974; Hazelworth and Weiselberg 1975

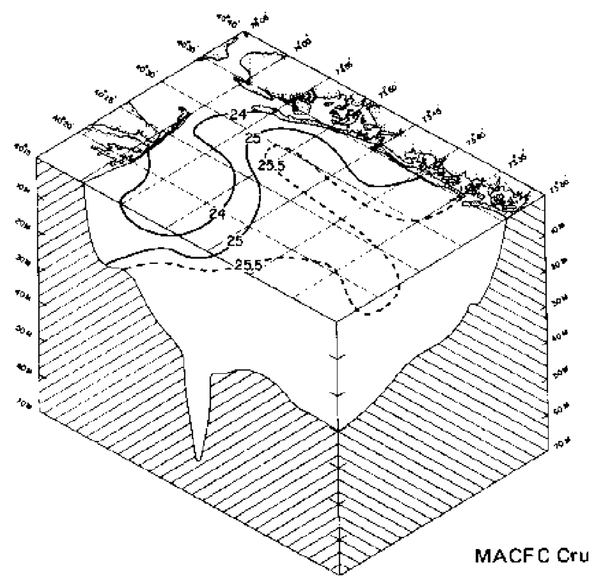
6-7 FEBRUARY 1969



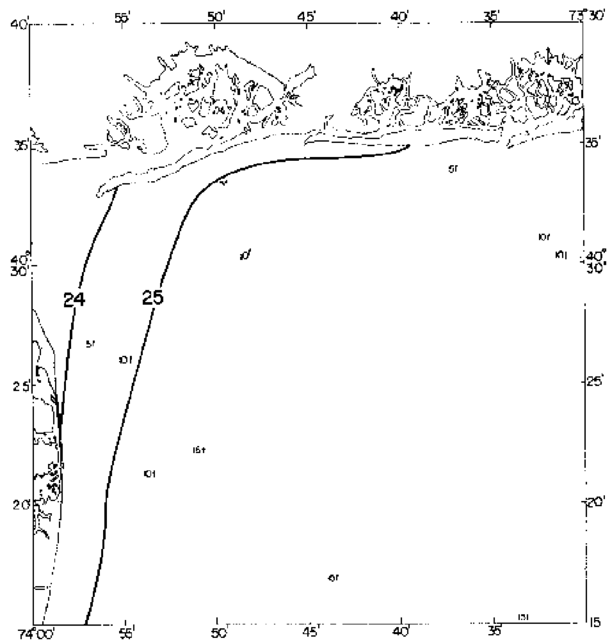
MACFC Cruise 1



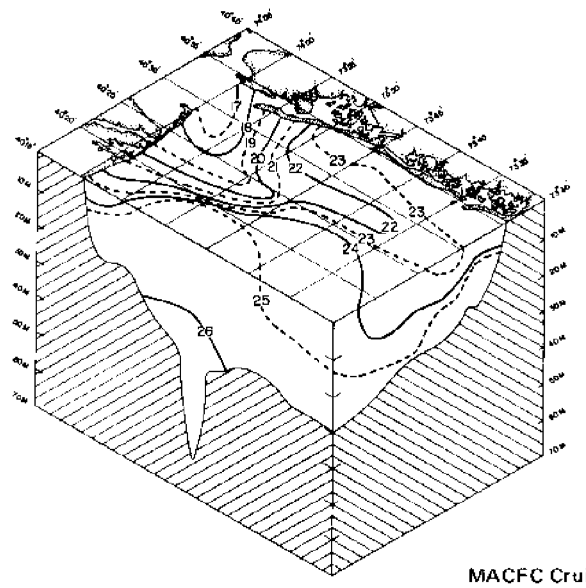
18-19 MARCH 1969



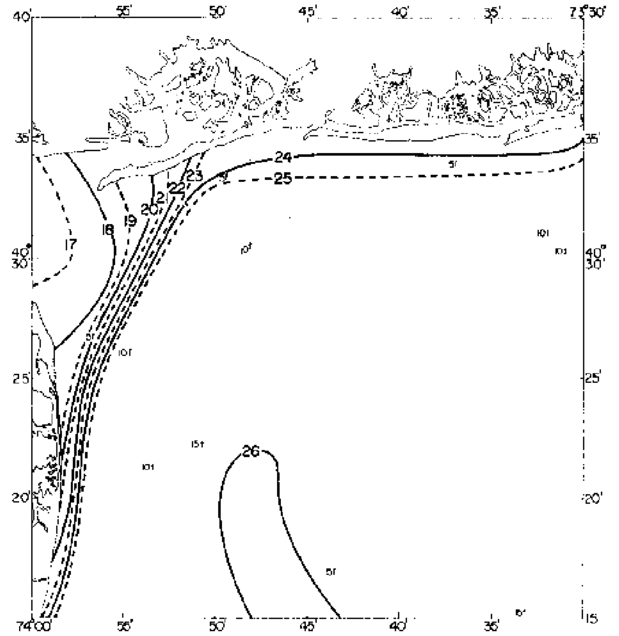
MACFC Cruise 2



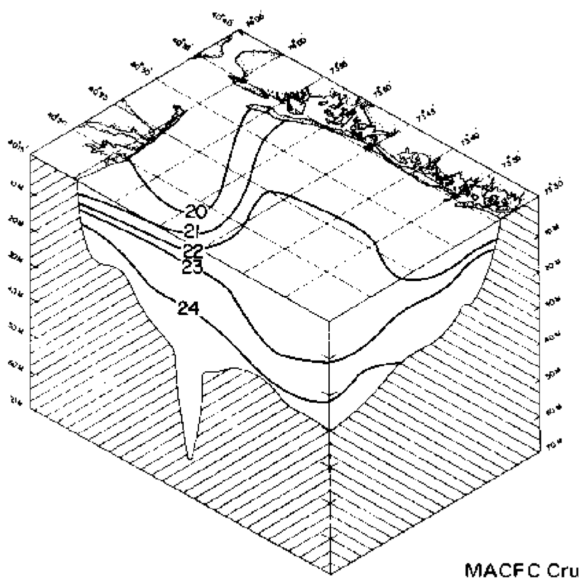
8-10 APRIL 1969



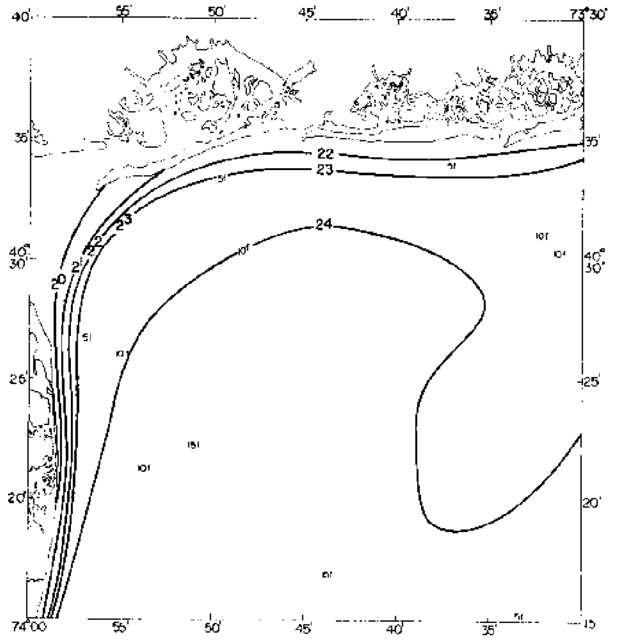
MACFC Cruise 3



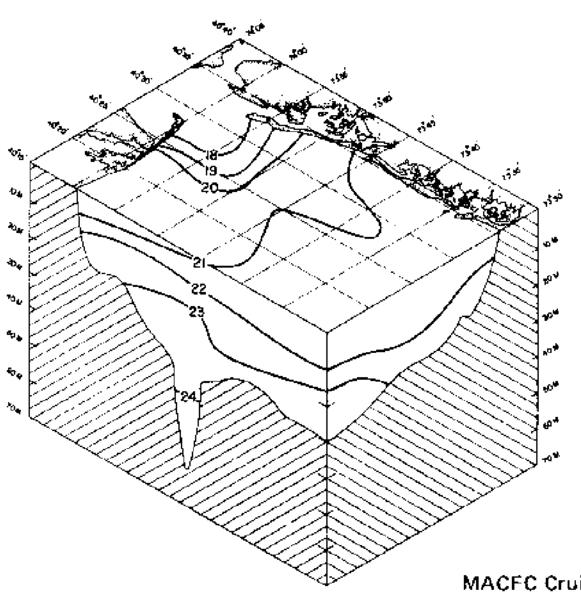
15-16 MAY 1969



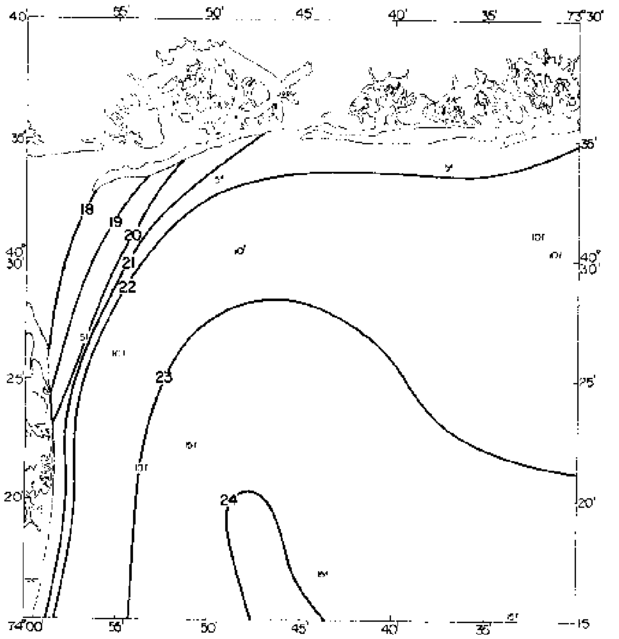
MACFC Cruise 5



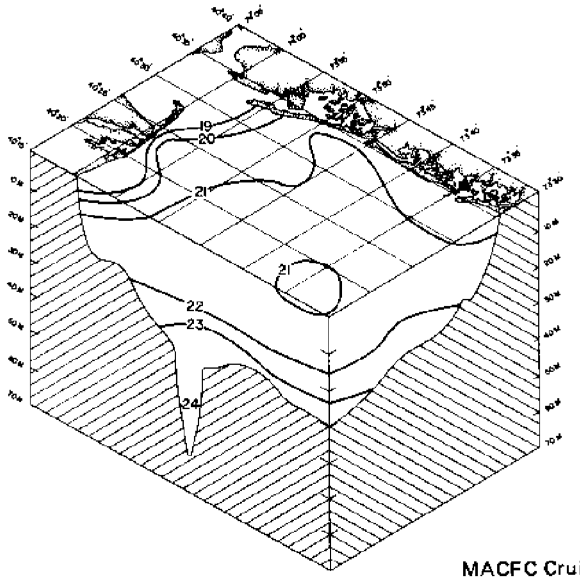
5-6 JUNE 1969



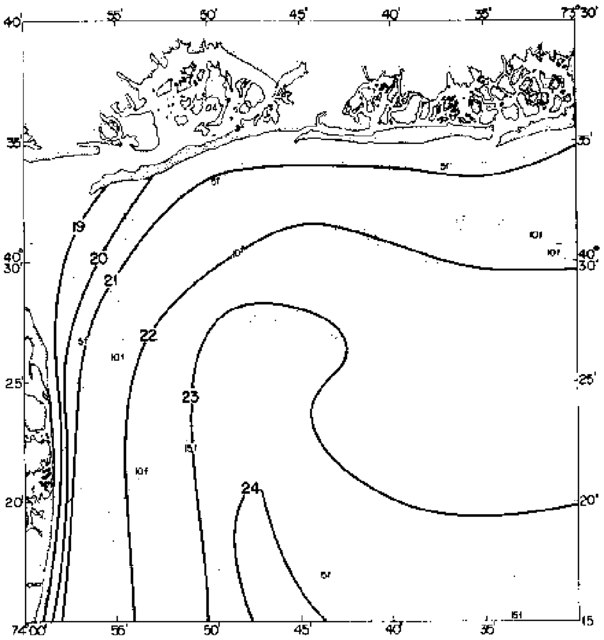
MACFC Cruise 6



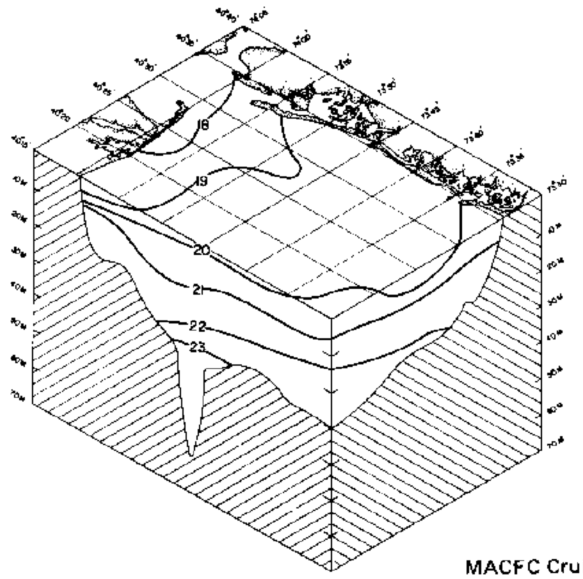
26-27 JUNE 1969



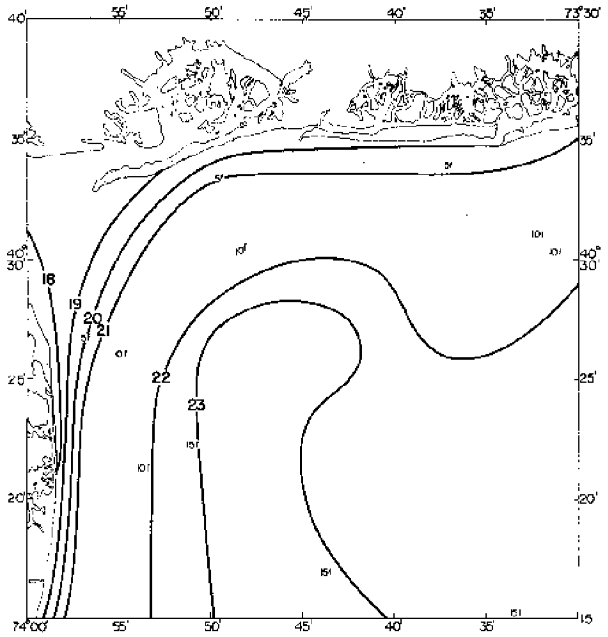
MACFC Cruise 7



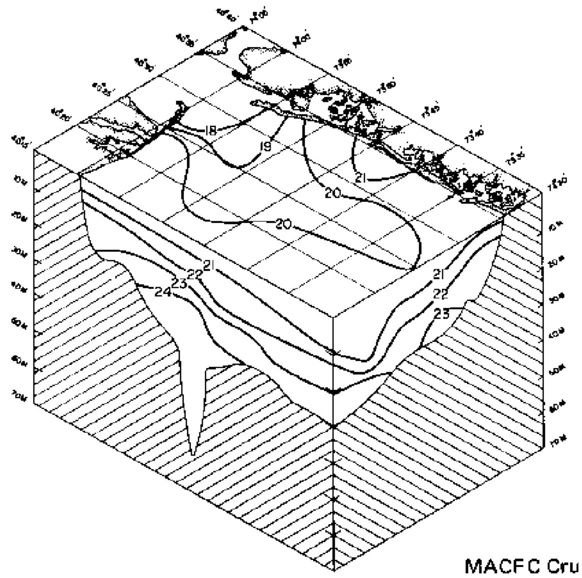
28 JULY-1 AUGUST 1969



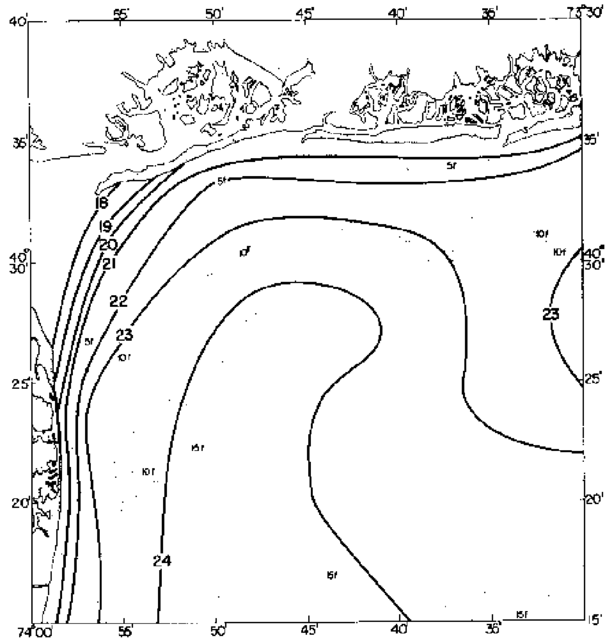
MACFC Cruise 8



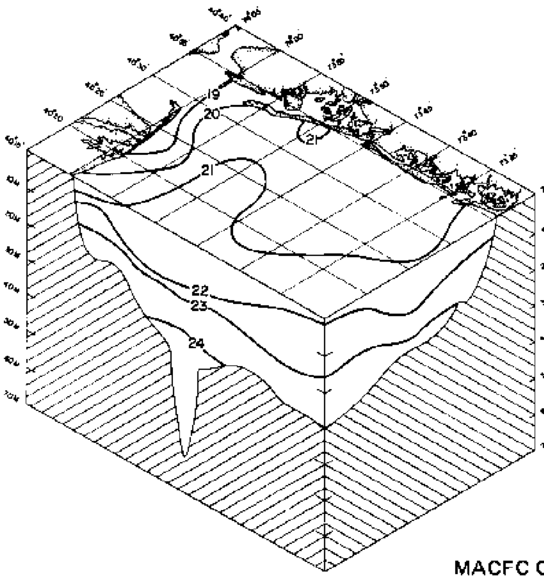
18-19 AUGUST 1969

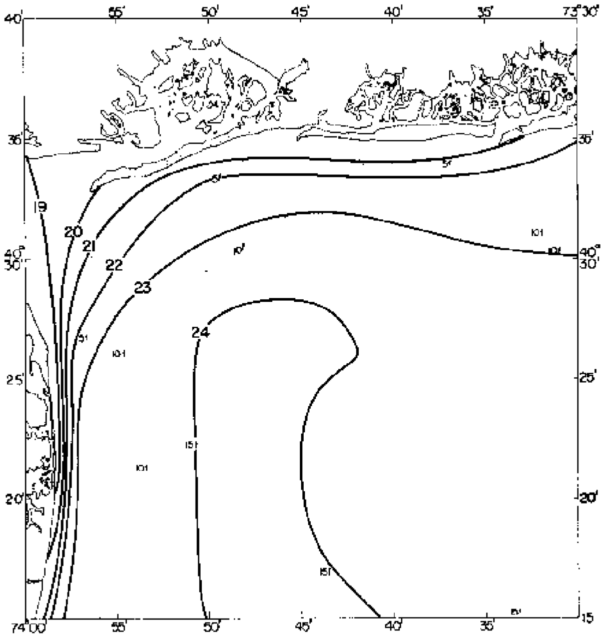


MACFC Cruise 9



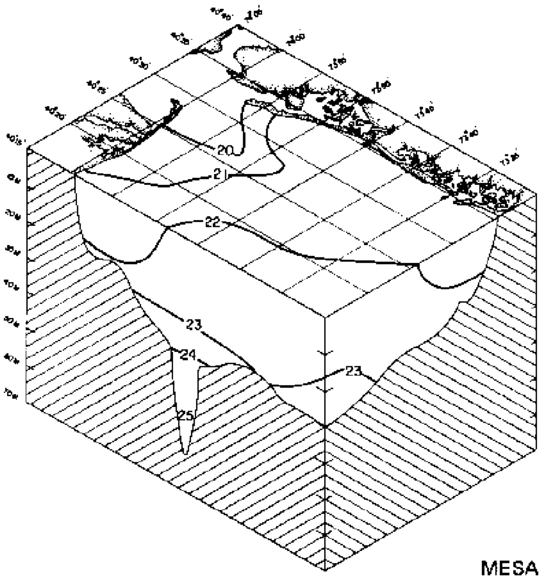
15-16 SEPTEMBER 1969

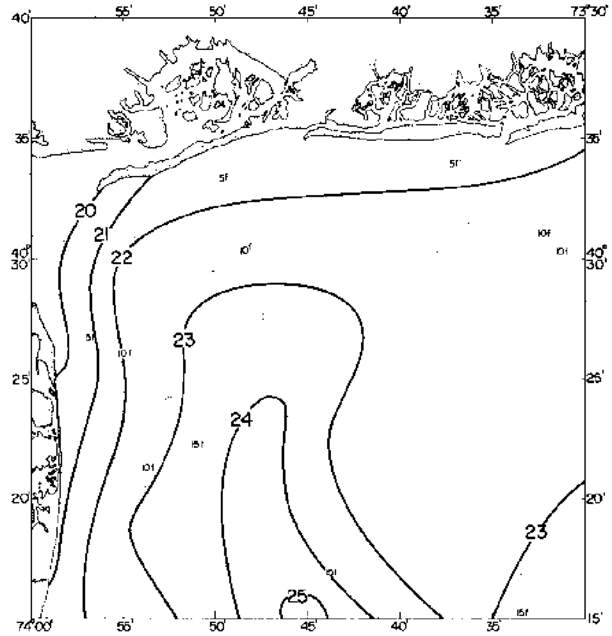




MACFC Cruise 10

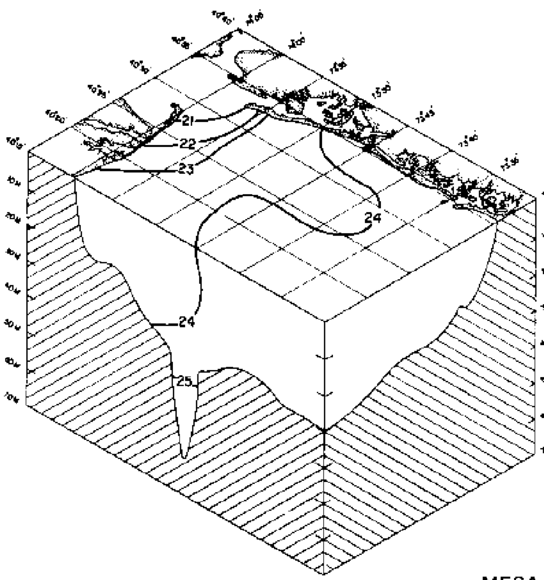
1-4 OCTOBER 1973

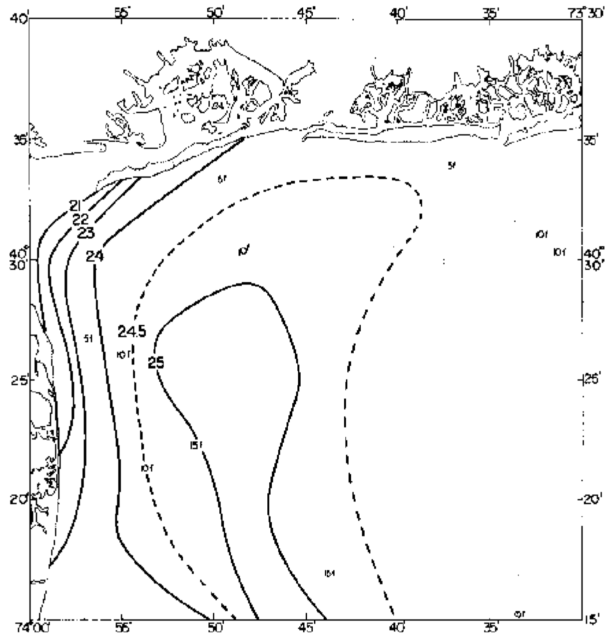




MESA Cruise 3

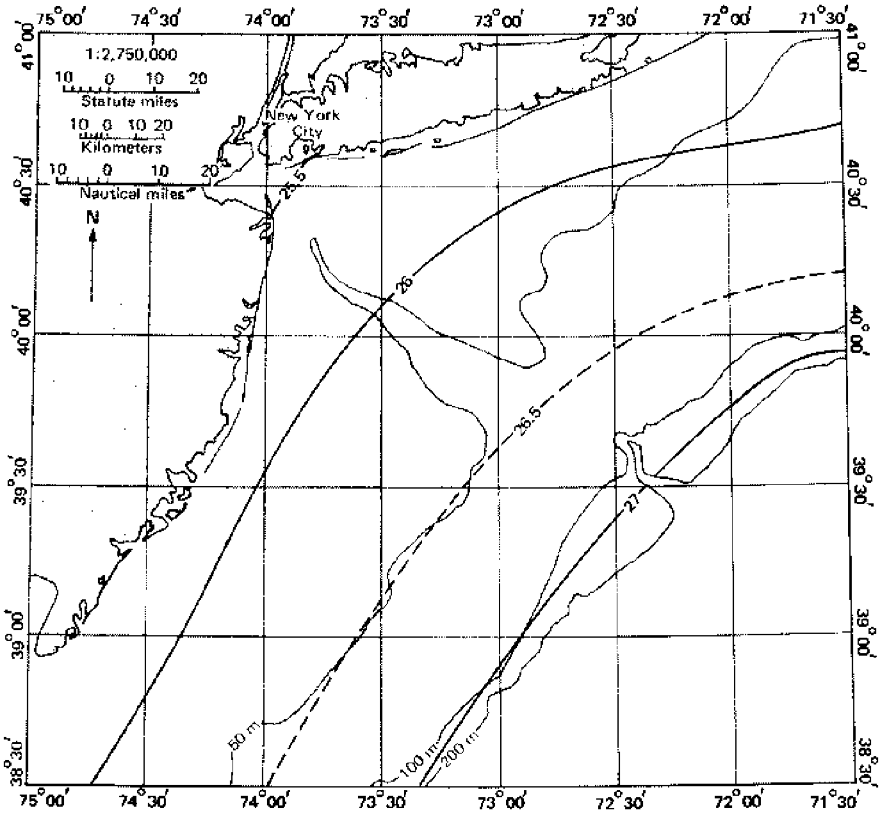
5-9 NOVEMBER 1973



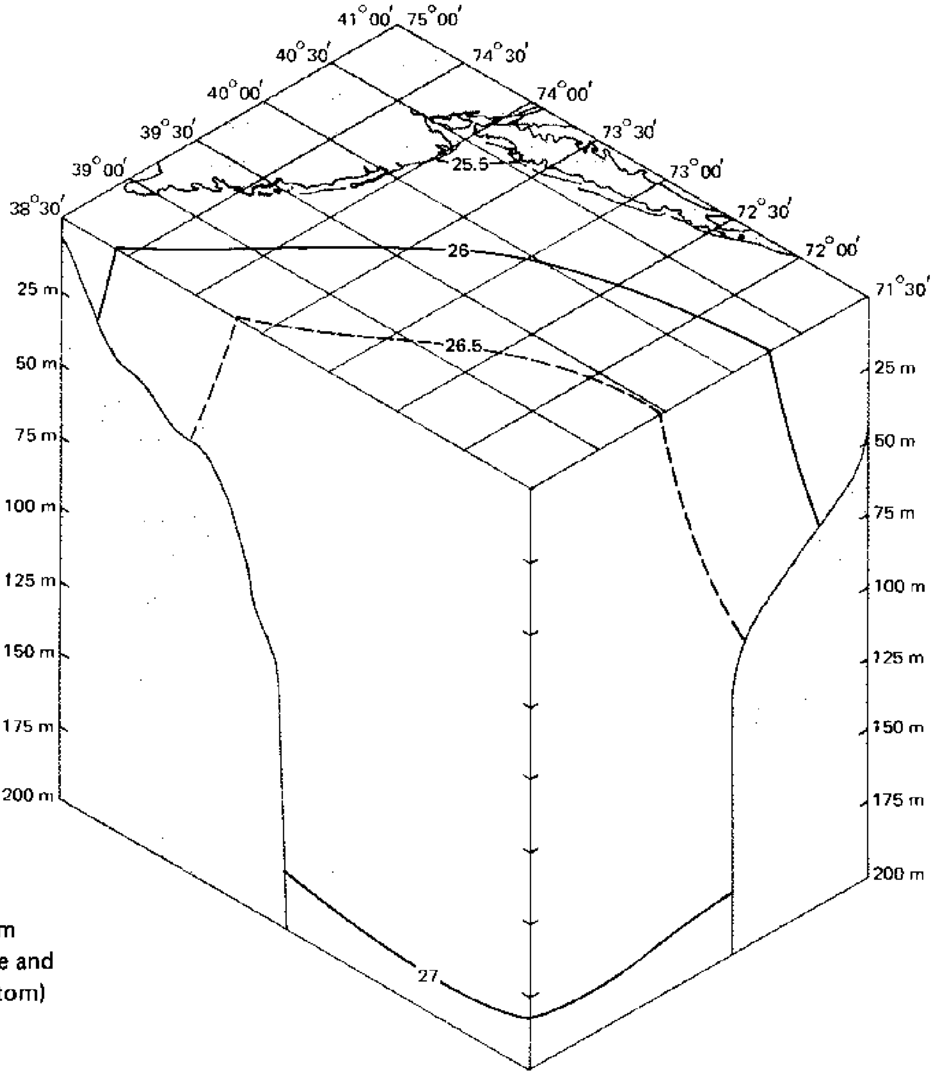


MESA Cruise 4

Map 10. Mean density distribution by season

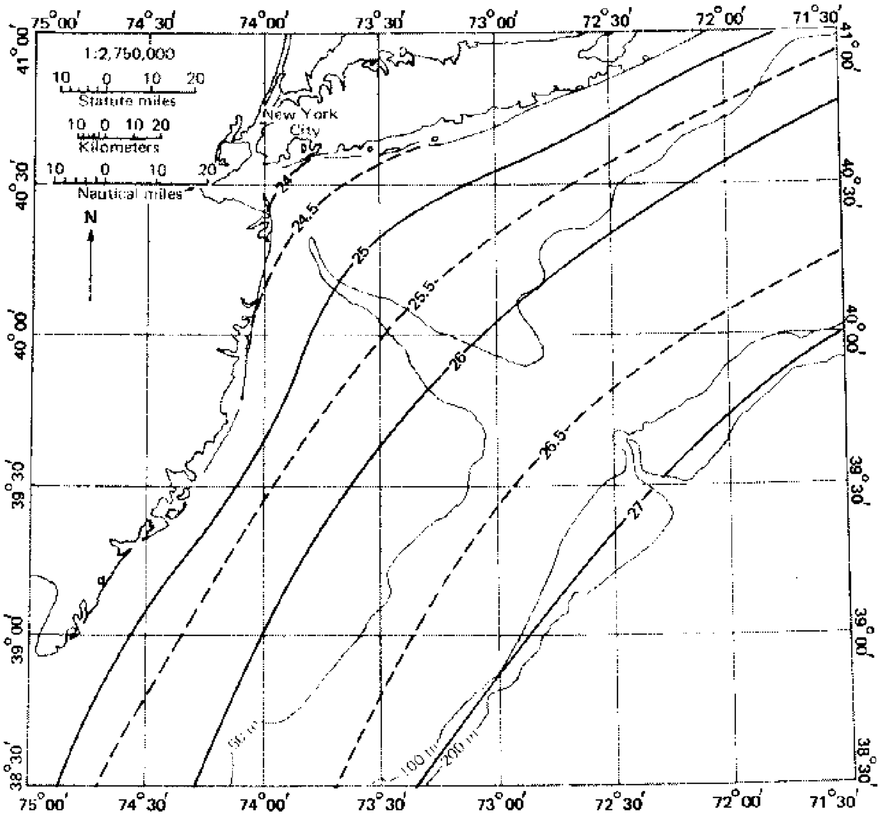


JANUARY-MARCH



Map pairs show bottom contours (top), surface and vertical contours (bottom)

Units are σ_T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Statute miles
Kilometers
Nautical miles
New York City
N
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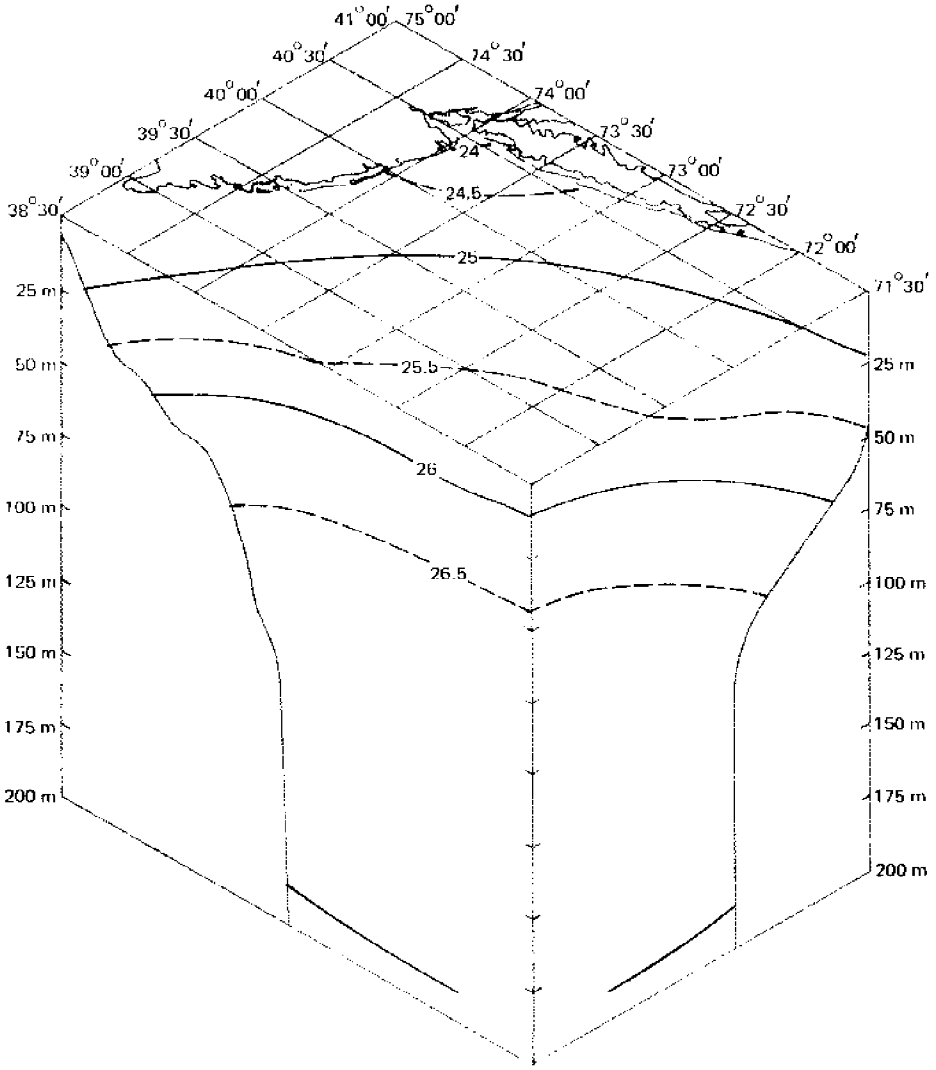
24.5
25
25.5
26
26.5
27
50 m
100 m
200 m

APRIL-MAY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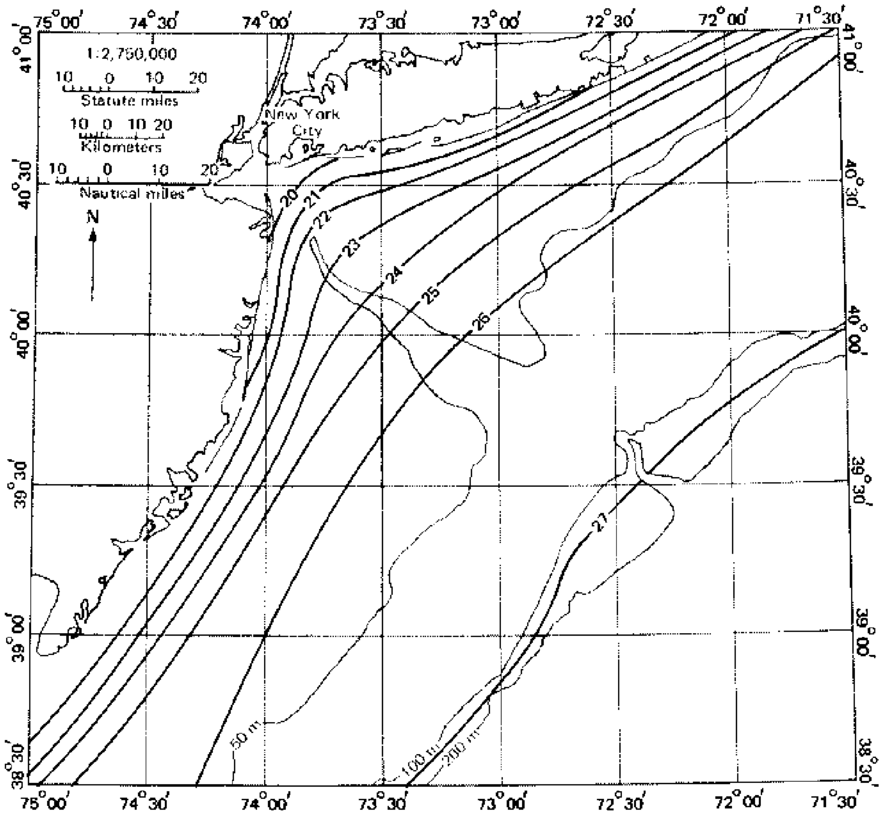
24.5
25
25.5
26
26.5
25 m
50 m
75 m
100 m
125 m
150 m
175 m
200 m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73°30′
73°00′
72°30′
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71°30′
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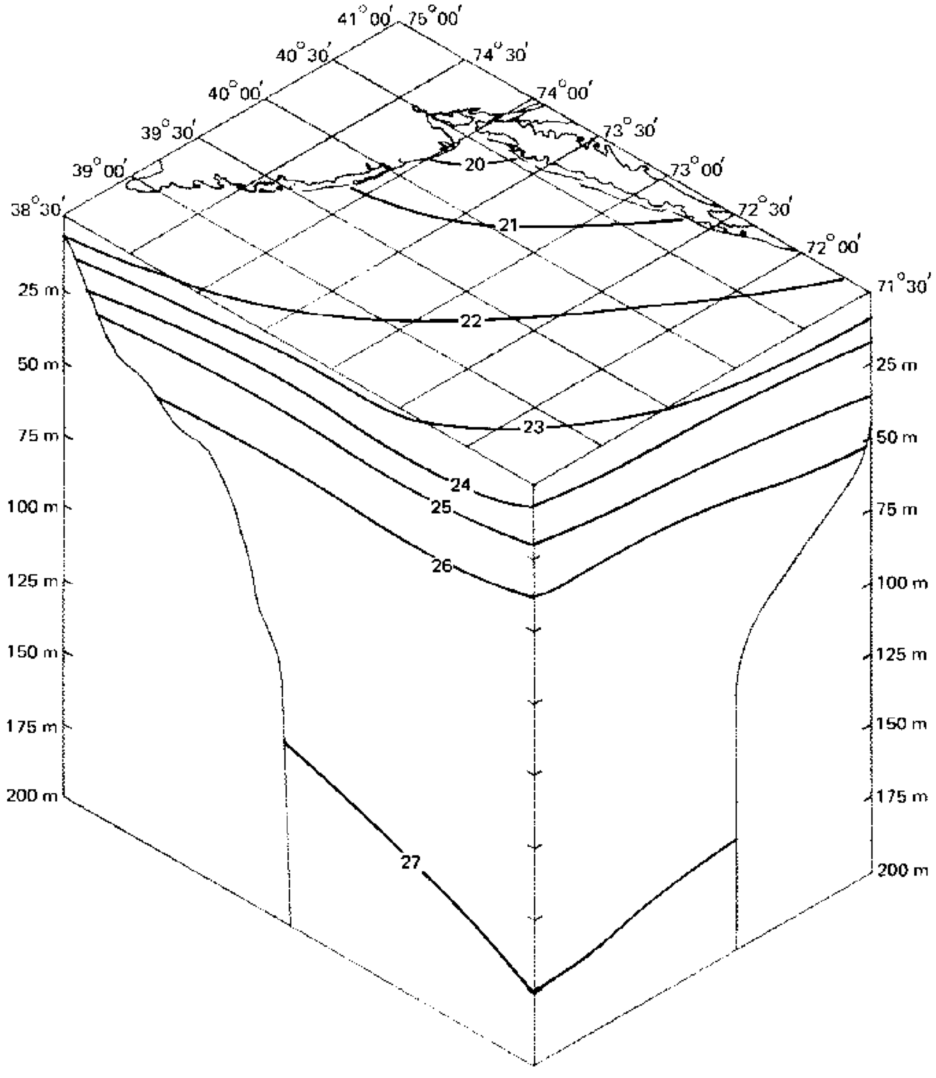
JUNE-AUGUST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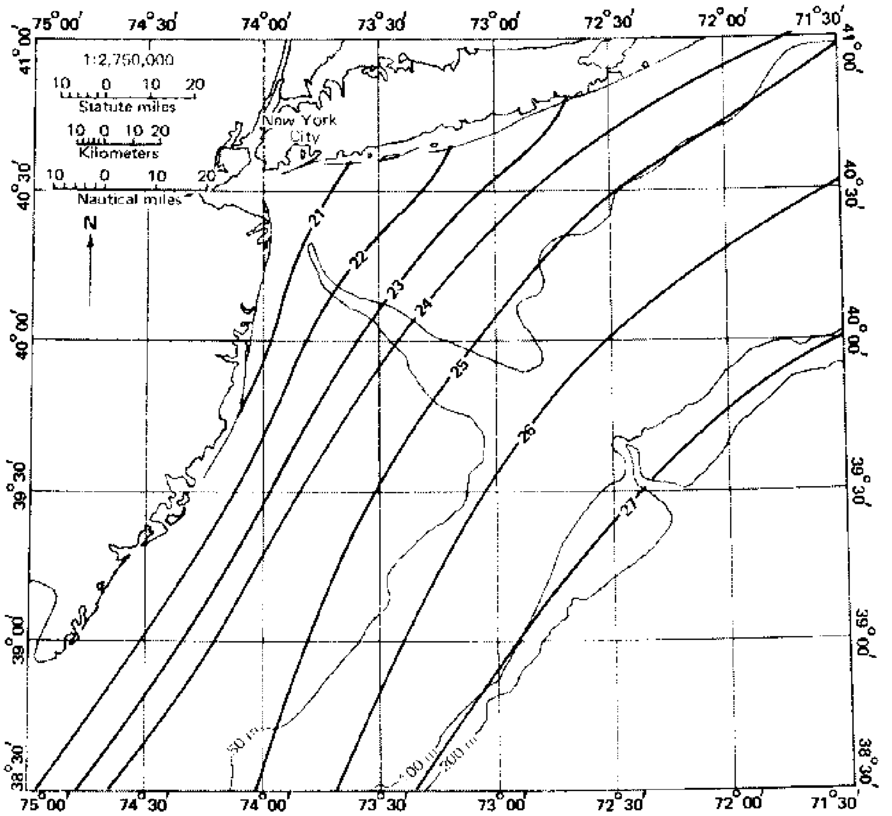
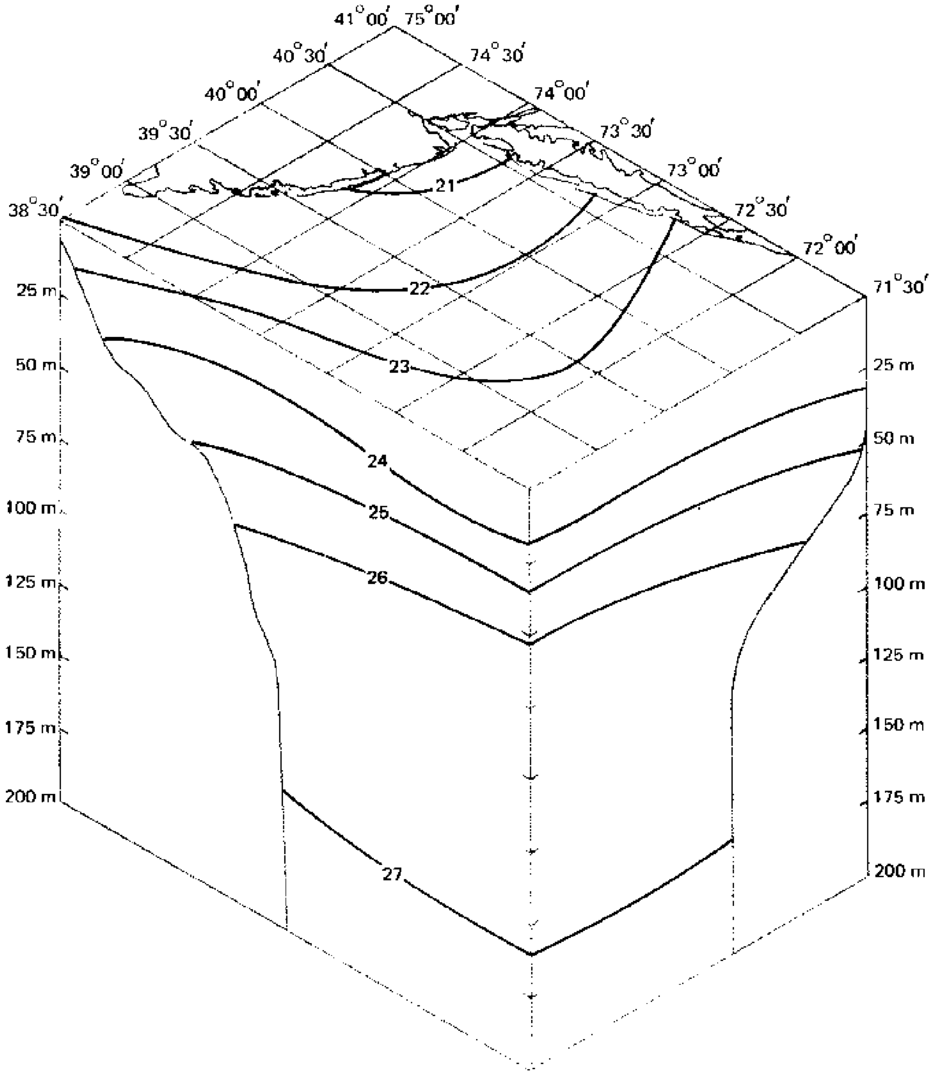
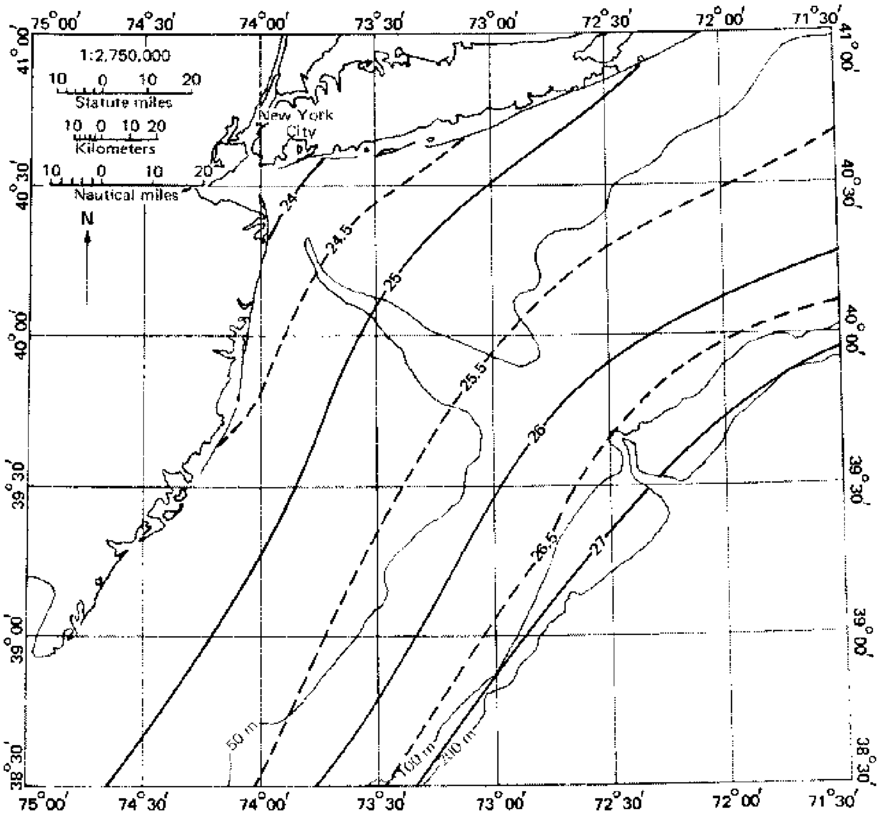


SEPTEMBER-OCTOBER





NOVEMBER-DECEMBER

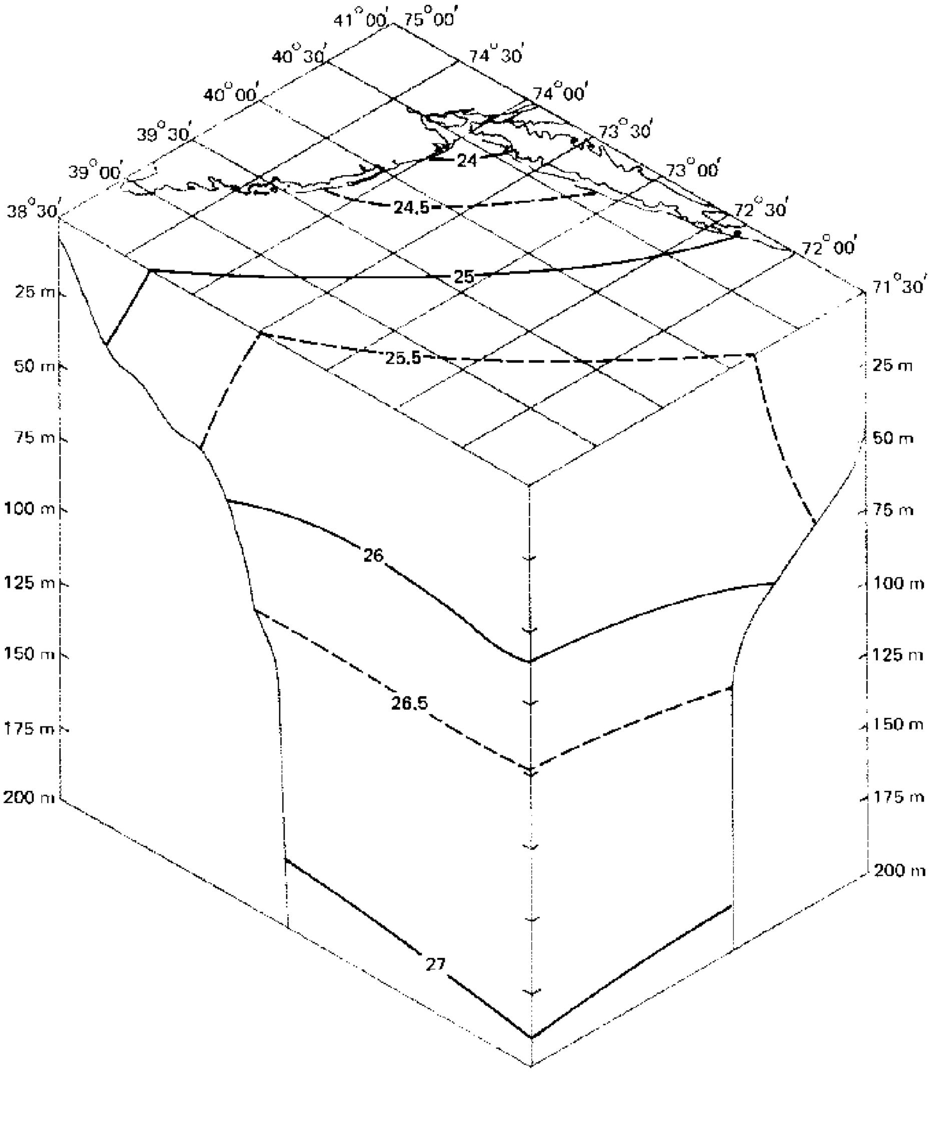


Figure 7. Winter temperature-salinity diagram and locator map

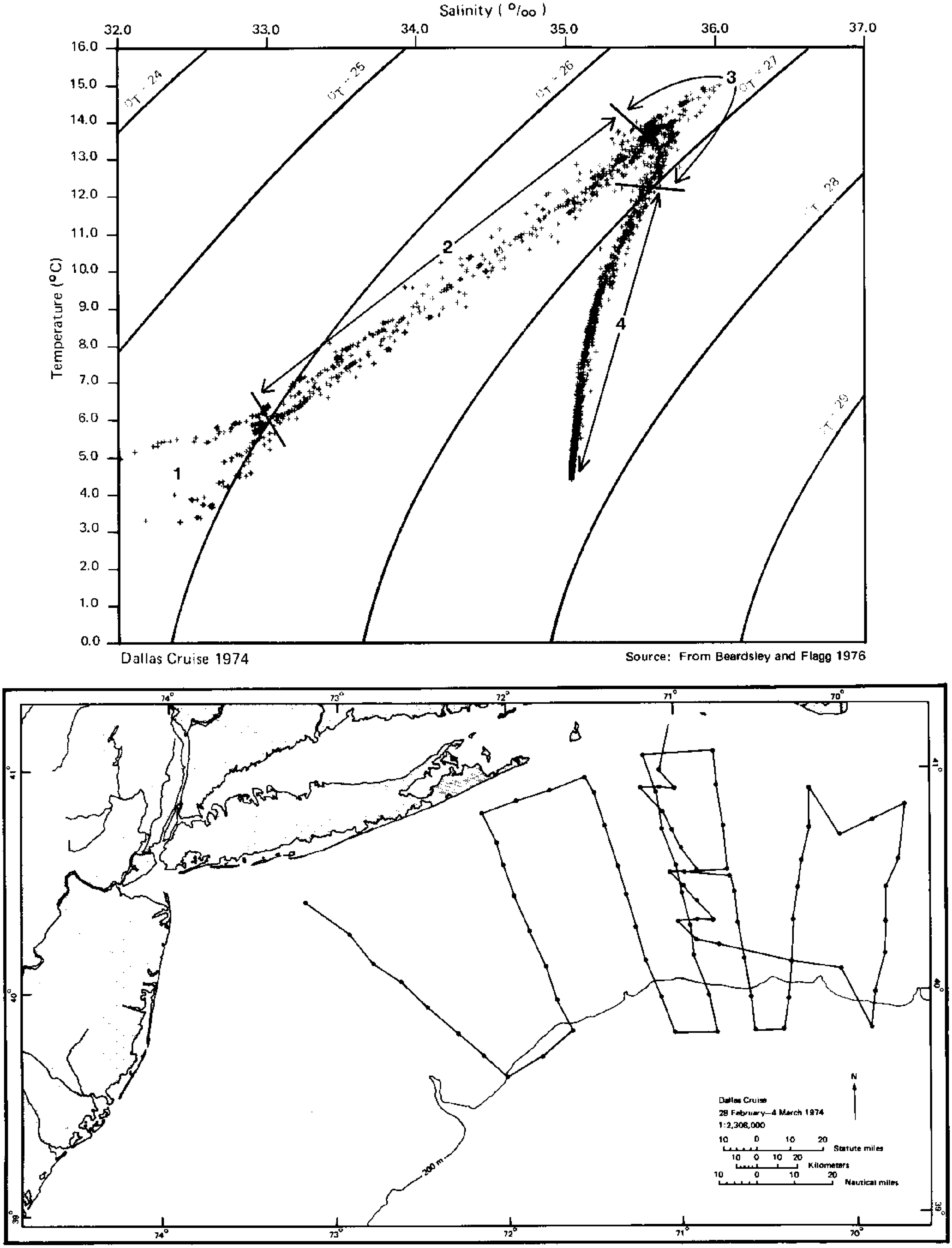
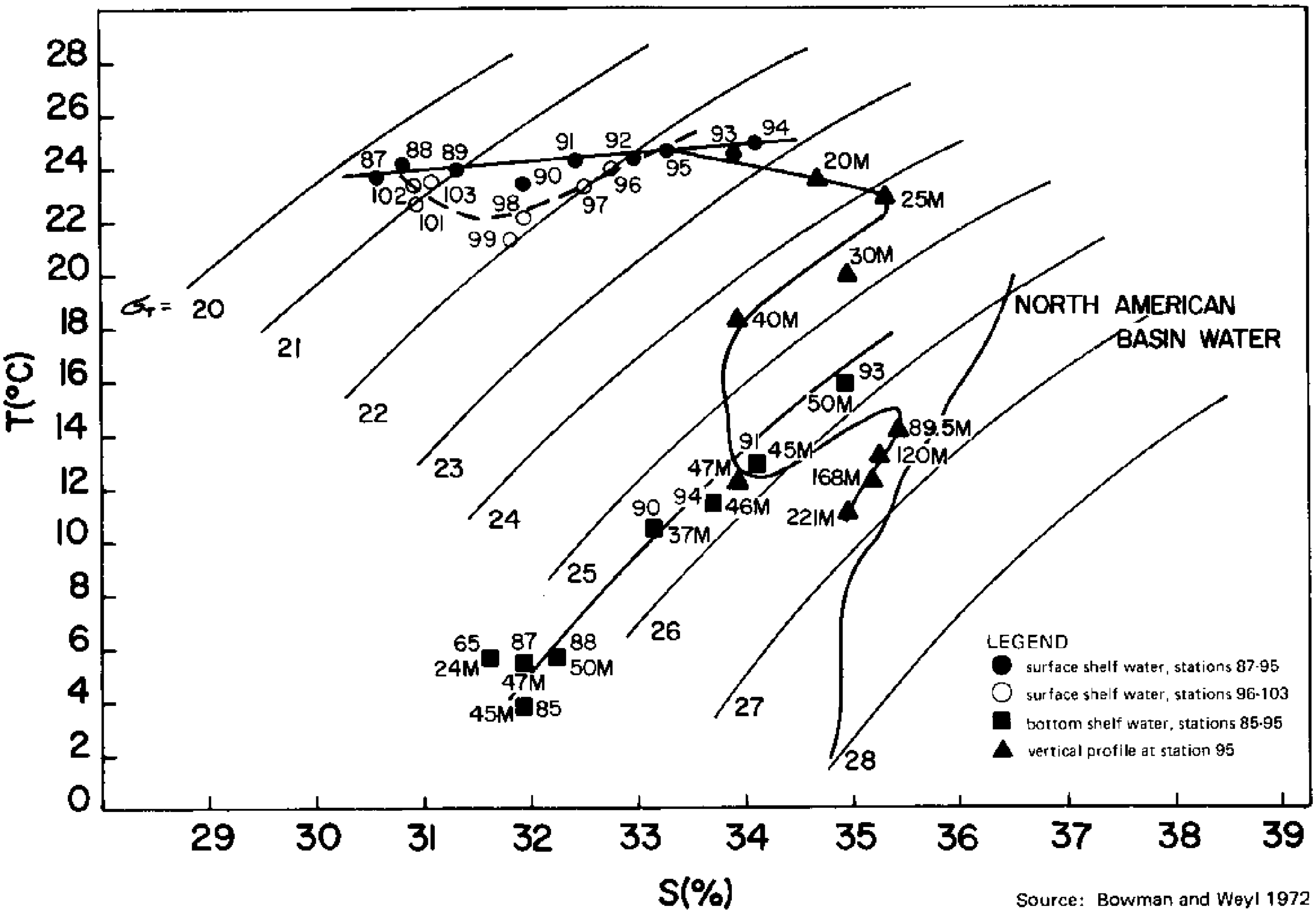
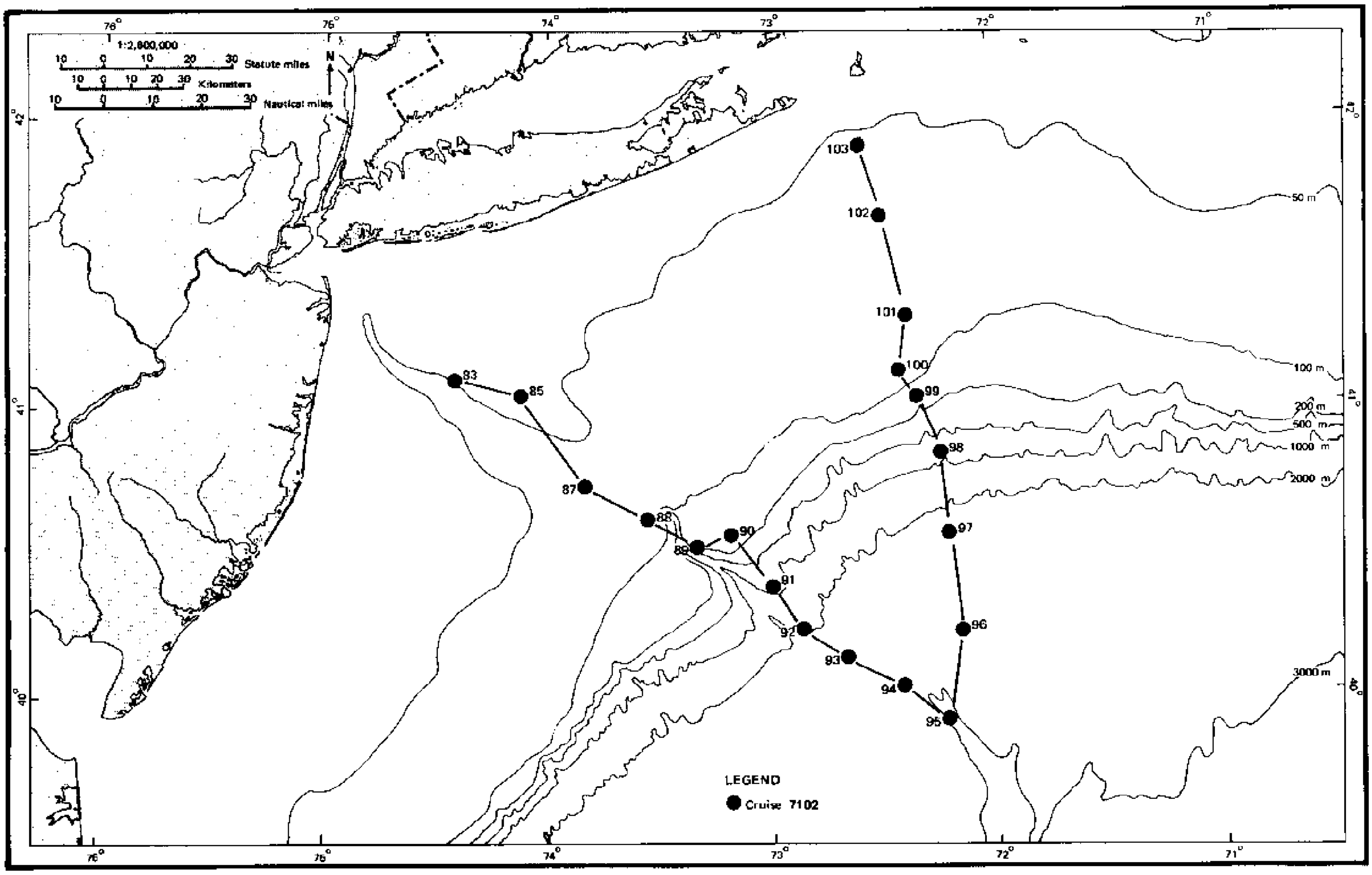


Figure 8. Summer temperature-salinity diagram and locator map



Source: Bowman and Weyl 1972



Averaging many years of hydrographic data obviously smooths out transient features like fronts, intrusions, meanders, and *eddies*. These have major influences on hydrographic properties, circulation, and the heat, salt, and *momentum budgets* of the Bight. Instabilities in the Gulf Stream flow lead to a number of significant phenomena, some directly influencing the oceanography of the Bight.

Slope or Oceanic Front

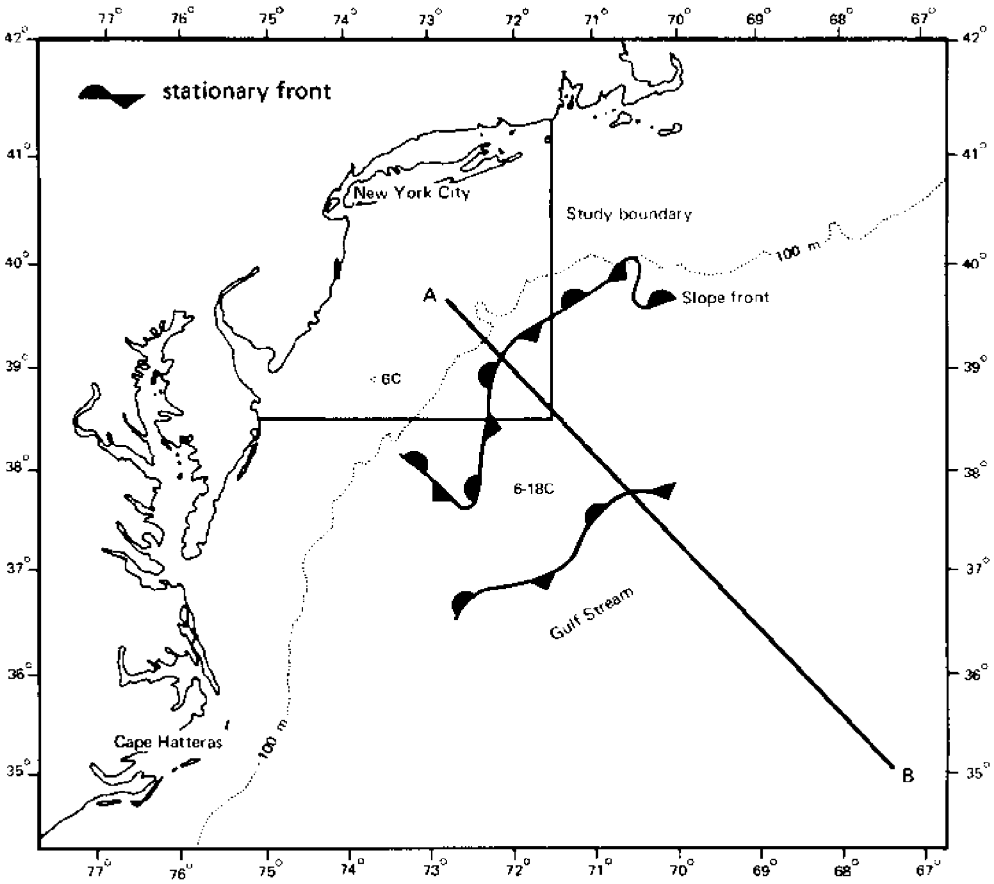
As described in the sections on temperature and salinity distribution, the boundary between shelf and slope waters in New York Bight is often delineated by a sharp front.

Although a salinity front may be observed during all seasons, a strong thermocline overlies the Bight in summer; any surface manifestation of a temperature front usually disappears between late June and September. The slope front in Map 11 forms the landward boundary over the temperature/salinity inversion, usually follows the bottom contours, and is often continuous throughout the Middle Atlantic Bight.

Figure 9 illustrates the relative location of various hydrographic features, including the developing seasonal thermocline and the "cool pool" on the shelf, the oceanic front, the temperature inversion over the slope, the Gulf Stream, and Sargasso Sea water. The Gulf Stream as shown in Map 11 is considerably closer to the continental shelf than normal and represents the intrusion of a major stream meander (US Naval Oceanographic Office 1970). An extrusion of the north wall at about 100 m has apparently penetrated onshore and may be important in maintaining the temperature maximum intersecting the shelf at 125 m.

Surface temperature contrasts across the winter front normally range from 2° to 4°C but have been

Map 11. Slope front and Gulf Stream, 31 March 1970



Source: US Naval Oceanographic Office 1970 observed as high as 7° or 8°C (US Naval Oceanographic Office 1969*a*, *b*). Salinity contrasts across the front are normally 1 to 2°/oo over a horizontal distance of 10 to 50 km and a vertical distance of 20 to 40 m.

Over short periods of time, portions of the front have been observed to move landward or seaward over ±75 km; meanders have been observed to propagate southwest along the frontal plane at velocities around 7.5 cm sec^{-1}, the opposite direction for Gulf Stream meanders (US Naval Oceanographic Office 1971*a*). Saunders (1971) found wavelike features on a late September front in the Gulf of Maine 65 km long and with a propagation velocity of 9.5 cm sec^{-1} to the west.

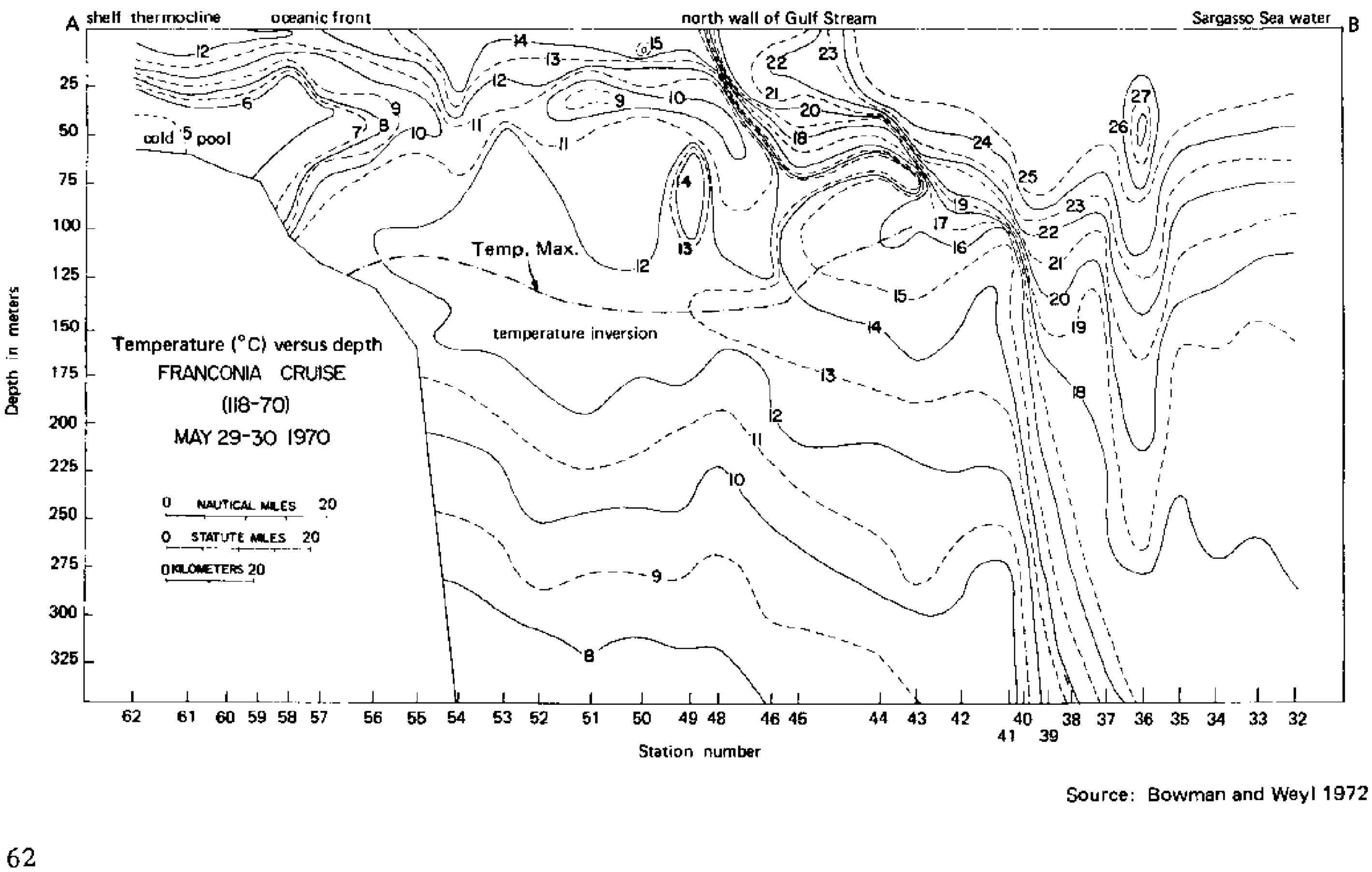
Another vertical temperature section across the shelf and slope of the Bight was taken May 1971 (Figure 10); the stations lie approximately along the transect AB in Map 11. This section exhibits a sharp slope front with temperature contrasts of 6°C at the surface. The front slopes shoreward and intersects the shelf around 100 to 150 m. The subsurface temperature maximum is clearly delineated, reaching from the surface under the front and intersecting the bottom at 150 m. Vernal warming has not commenced and temperatures on the outer shelf are almost isothermal around 6°C.

Surface temperatures along a section taken 7 August 1971 (Figure 10) are close to the annual maximum of 24° to 25°C. However, offshore and below the summer thermocline, a temperature inversion persists around 150 m, underlies an extrusion of 10°C shelf water at 60 m, and lies in the transitional zone between the shelf and slope thermoclines. Strong temperature contrasts around 4°C can be seen at the shelf break, separating cold bottom shelf water from subsurface slope water.

In autumn, surface cooling over the shelf is more rapid than over the slope, reestablishing the surface frontal zone with temperature contrasts of 4°C, illustrated in the 2 October 1971 section (Figure 10). The seaward slope of the isotherms in the thermocline over the shelf break produces strong horizontal gradients; temperatures at 60 m increase from 7° to 15°C over 25 km. An intrusion of surface slope water penetrates the shelf thermocline at 25 m, and a local temperature minimum at 150 m is from the extrusion of shelf cold pool water over the break.

The 31 October 1971 section (Figure 10) shows continuing deepening in the shelf isothermal surface layer during late fall and the general structure of the permanent thermocline over the slope where the surface layer is 40 m thick. At the edge of the shelf can be seen a front with a temperature contrast of

Figure 9. Vertical temperature profile along line parallel to A-B in Map 11



Source: Bowman and Weyl 1972

about 2°C. A frontal zone at the shelf break also exists below the thermocline and separates cold pool bottom water (10°C) from intermediate slope water (14° to 15°C).

Little is known about *velocity shears* across these fronts. Persistent oceanic thermal fronts have also been observed in the Sargasso Sea, meandering generally east-west (Voorhis and Hersey 1964; Voorhis 1969). These wedgelike fronts gradually incline downward 1° beneath the surface toward the south, penetrating to a depth of 50 m in 10 km. Strong currents 70 to 80 cm sec^{-1} were found flowing as a narrow jet to the east along the warm frontal edge. On the cold water (north) side of the front the measured currents were relatively low (10 to 30 cm sec^{-1}) and flowed in the opposite direction (westward). Some places evidenced convergence toward the frontal edge.

Based on geostrophic assumptions, Beardsley and Flagg (1976) calculated an expected shear of 10 cm sec^{-1} across a March, New England slope front, with inshore water flowing westward parallel to the frontal edge. However, a few current meter records did not confirm this predicted shear, although admittedly it is very difficult to measure.

In summary, the appearance of the surface temperature front in autumn and its disappearance in spring results from the different rates of heat loss and gain between shelf and slope waters. Because of less heat storage, shelf waters cool and warm faster than slope waters at the same depth. In winter, near the shelf break heat and water loss cause surface slope waters to sink and flow under surface shelf waters, forming the temperature/salinity inversion. In summer this inversion appears to be maintained by intrusions of offshore slope water spreading shoreward under the slope thermocline along isopycnal surfaces.

Studies of the phenomena are complicated by large variability, by our lack of knowledge of important small-scale processes, and by intrusions of Gulf Stream meanders and eddies into the region. The stability of the fronts is necessarily dependent on relative inclinations (0.1° to 0.2°) and the inclination of the shelf break (0.1° to 0.4°). Clearly the fronts are the end products of a large-scale wind field and seasonal heating and cooling of the water column. The actual formation of the fronts, however, must depend on small-scale processes at the frontal edge (Voorhis 1969; Garvine 1974*a*). Understanding these processes is the key to *frontogenesis* in the Bight region.

Gulf Stream Meanders

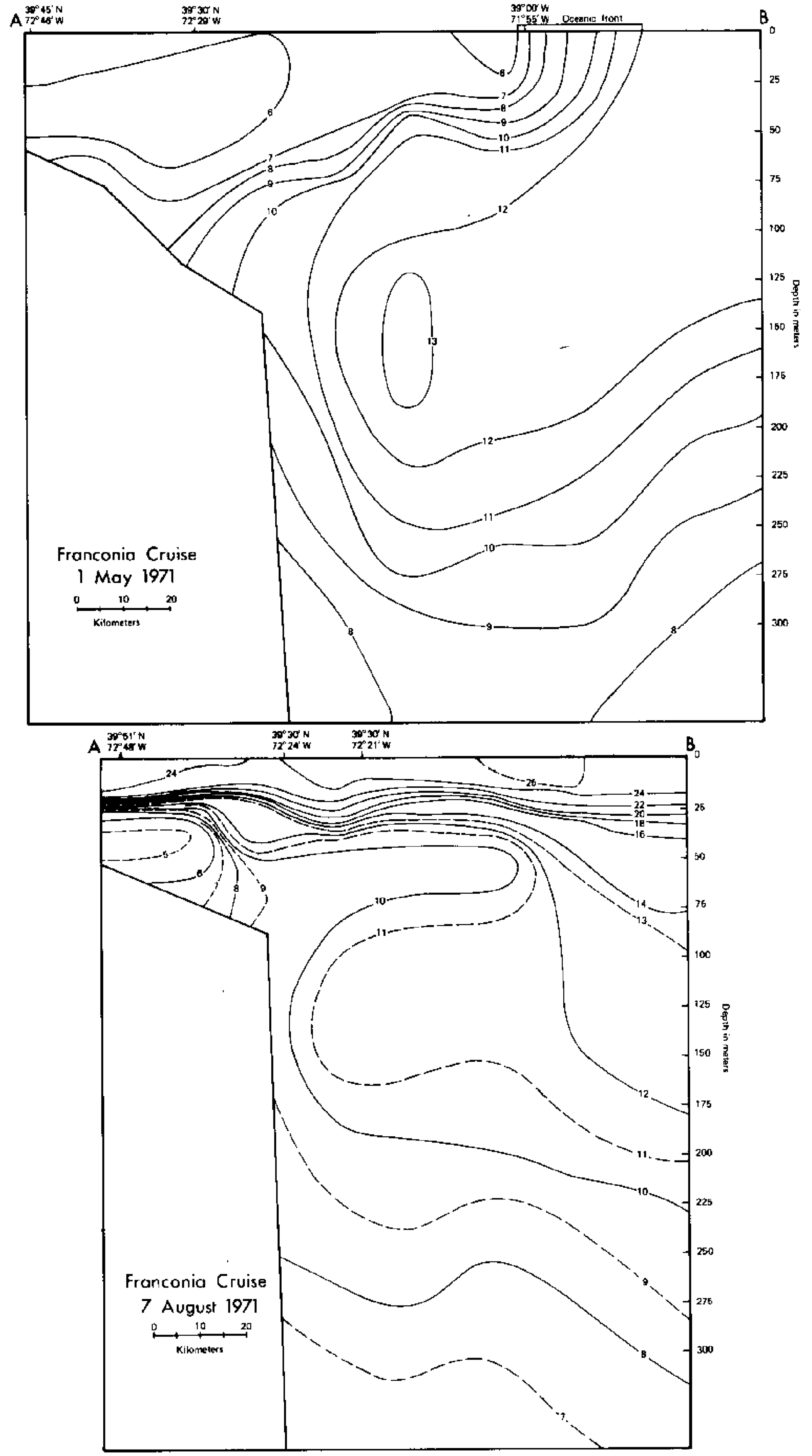
Gulf Stream meanders and the consequent spawning of anticyclonic (clockwise) eddies inshore and cyclonic (counterclockwise) eddies into the Sargasso Sea have been known for many years (Iselin 1936). The sinuous meandering of the Gulf Stream between Cape Hatteras and the Grand Banks has been studied in ship and aircraft surveys (Fuglister and Worthington 1951; Fuglister 1963; Fuglister and Voorhis 1965; Wilkerson and Noble 1970; Gotthardt and Potocsky 1974). Small changes in the slope of bottom topography have been suggested as responsible for steering the Gulf Stream axis (Warren 1963; Niiler and Robinson 1967; Robinson and Niiler 1967). Theoretical models incorporating both topographic steering and hydrodynamic instabilities have been developed to describe (with some success) the mean position and *envelope*, and the wavelengths and propagation velocities of meanders (Orlanski 1969; Robinson 1970). Hansen (1970) observed the paths of many Gulf Stream meanders and concluded that none of the existing theories satisfactorily accounts for all major features of meandering.

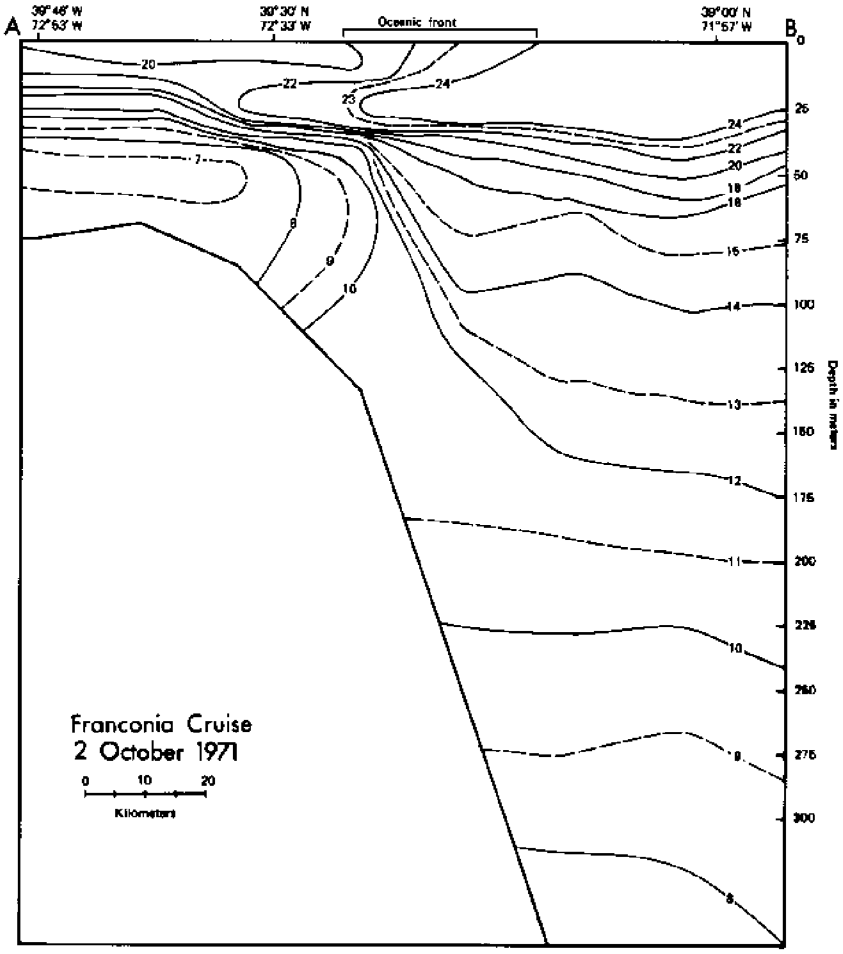
Only over the last few years has it been possible to survey these meanders and their movements intensively and systematically, largely because of expendable bathythermograph (XBT), aircraft XBT (AXBT), and very high resolution infrared radiometers abroad satellites. Although much of the Middle Atlantic Bight is covered with clouds for long periods—for example, the sea surface was visible approximately once each 3.3 days between December 1973 and March 1974 (US Naval Oceanographic Office 1974*a*)—it has been possible to delineate and follow short-term movements of the slope front, anticyclonic eddies, and the northern edge of the Gulf Stream.

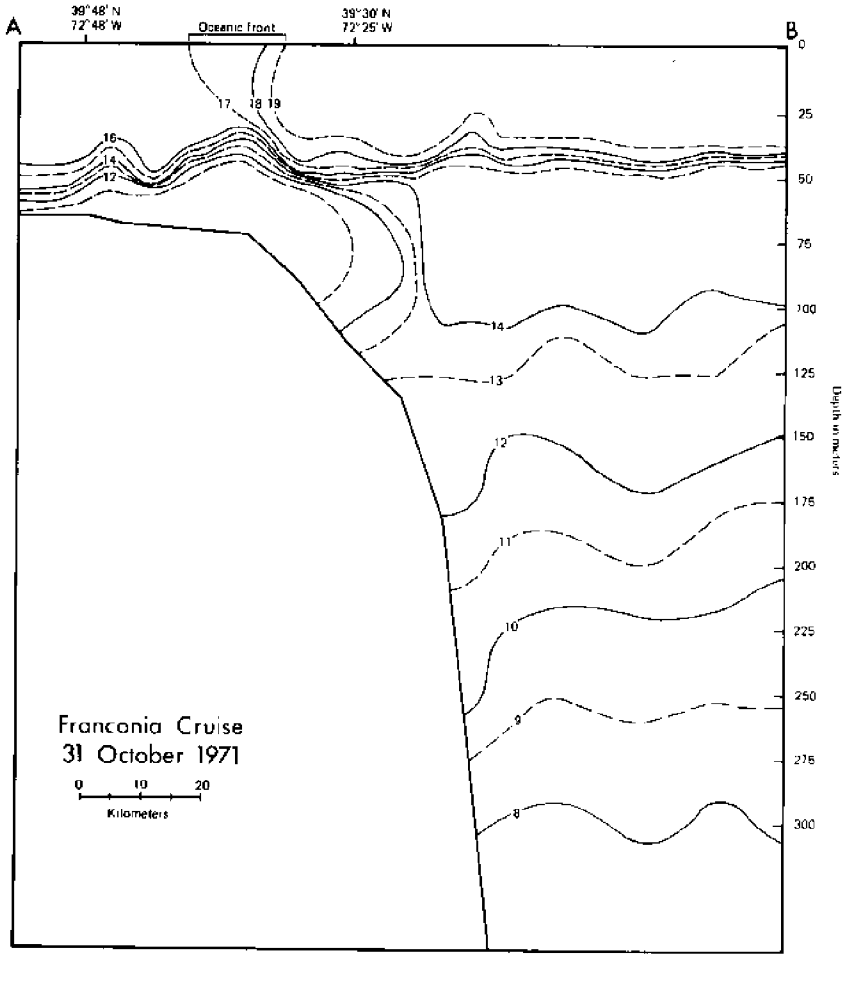
Between 1970 and 1973, 108 transits across the Gulf Stream were made by Naval Oceanographic Office personnel aboard commercial ships. Statistics gathered from these cruises show that the position of the Gulf Stream north wall varied 166 km along a line (AB in Map 11) during this time, although for 95% of the surveys the movement was confined to 95 km. The most common cause of this lateral movement was meandering (US Naval Oceanographic Office 1974*e*).

A rare meander penetrated shelf waters of the Bight during late July and early August 1973 (US Naval Oceanographic Office 1973*b*, 1974*e*; Gotthardt and Potocsky 1974). Bowman and Duedall (1975)

Figure 10. Vertical temperature profiles, Franconia cruises 1971




A
39°48' W
72°63' W
39°30' N
72°33' W
Oceanic front
39°00' N
71°57' W
B
0
25
50
75
100
125
150
175
200
225
250
275
300
Depth in meters
20
22
24
23
7
8
9
10
18
16
14
13
12
11
Franconia Cruise
2 October 1971
0
10
20
Kilometers


A
39°48' N
72°48' W
Oceanic front
39°30' N
72°25' W
B
0
25
50
75
100
125
150
175
200
225
250
275
300
Depth in meters
17
18
19
16
14
12
13
11
10
9
8
Franconia Cruise
31 October 1971
0
10
20
Kilometers

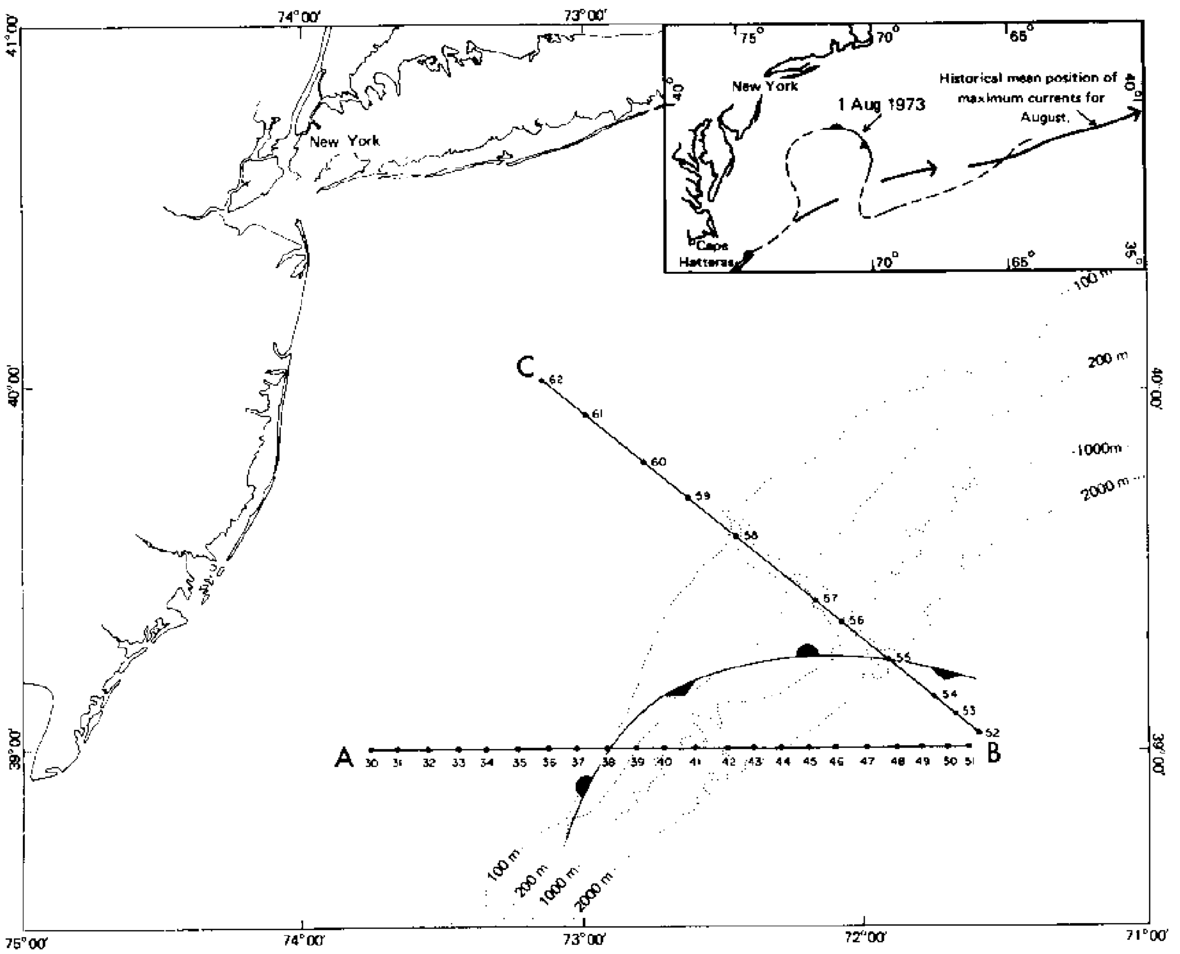
made two transects across the head of this meander (Map 12). The vertical temperature section in Figure 11 illustrates how shallow overrunning of the northern surface edge of the Gulf Stream has penetrated slope and shelf waters over the 100 m isobath. Large horizontal gradients of 10°C in 2 km were recorded over the shelf break.

The salinity section shows a very sharp vertical salinity front with a surface contrast around 4.5°/oo separating shelf water from Gulf Stream water between stations 38 and 39. Shelf salinities range between 31 and 33°/oo, whereas values inside the stream reach over 36.5°/oo.

The pycnocline established over the shelf by the summer thermocline increases in depth over the slope and into Gulf Stream water to the right of the density section in Figure 11. No evidence of sharp horizontal density gradients is found across these temperature and salinity discontinuities. A second set of vertical sections taken across the meander is shown in Figure 12. The stream has moved offshore and the northern edge (15°C isotherm at 200 m) lies at the right-hand edge of Figure 12. Further evidence of the edge of the stream is shown in the salinity section by the transition from horizontal to vertical isohalines over the slope. Horizontal gradients on this section are much weaker and suggest the northern edge lies at a shallow angle to the line BC. Again no density discontinuity exists at the Gulf Stream in Figure 12.

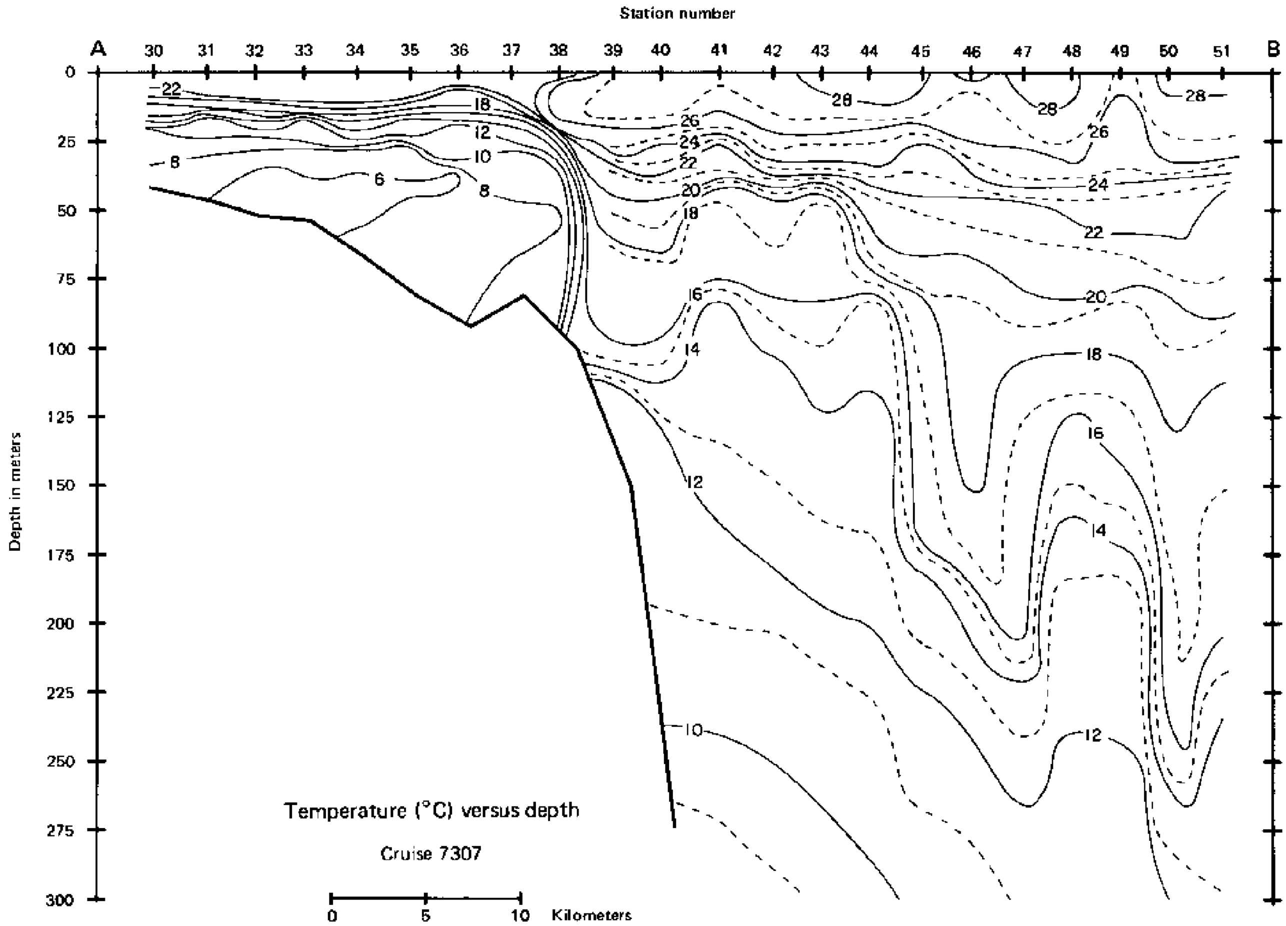
The event is summarized in a T-S diagram (Figure 13). Surface slope water, with temperatures and salinities intermediate between shelf and Gulf Stream surface waters, has been displaced by the meander. Bottom shelf and slope water masses are shown; a vertical profile at station 52, and the T-S characteristics of North American Basin water (Wright and Worthington 1970) are included for comparison.

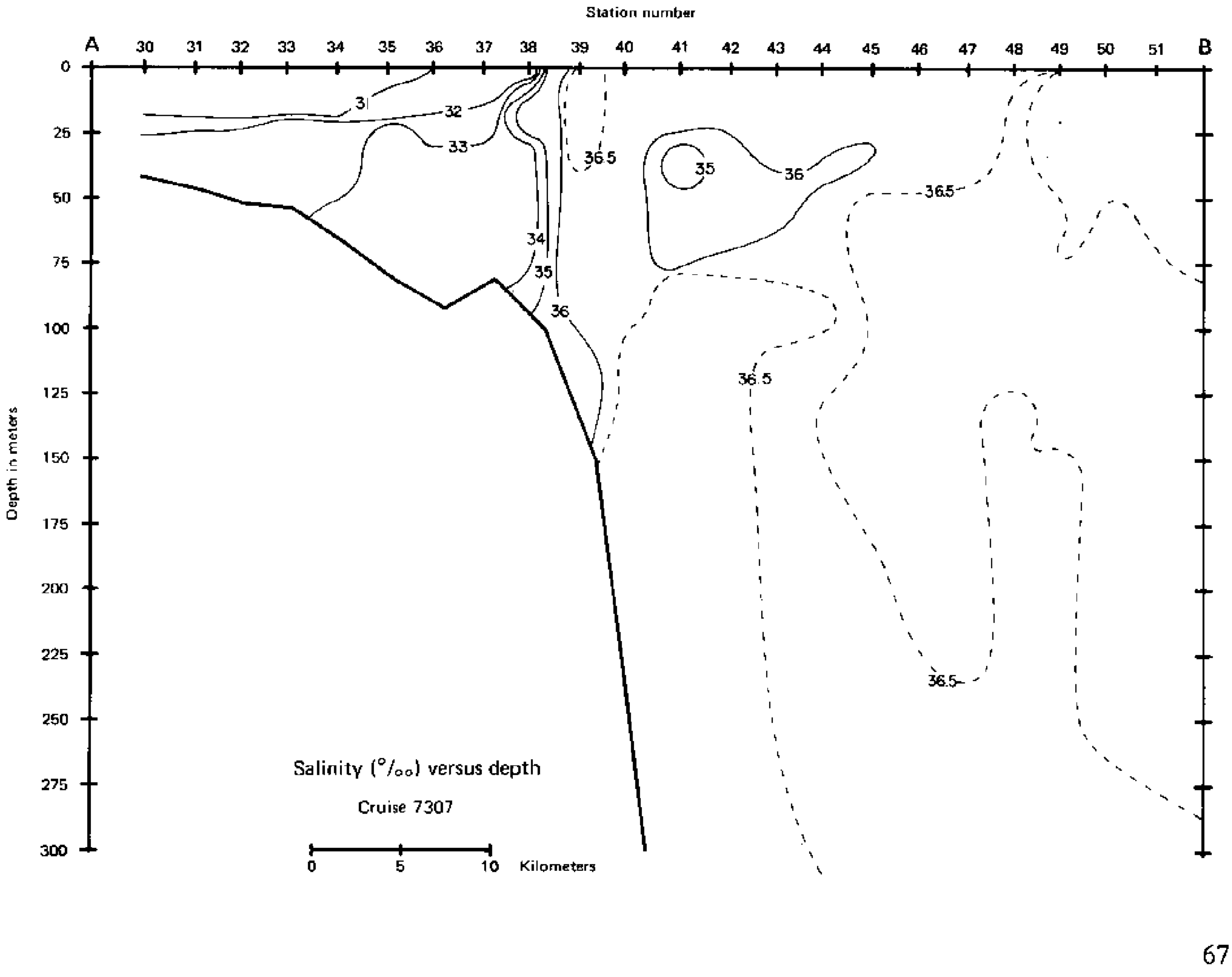
Map 12. 1973 Gulf Stream meander locator

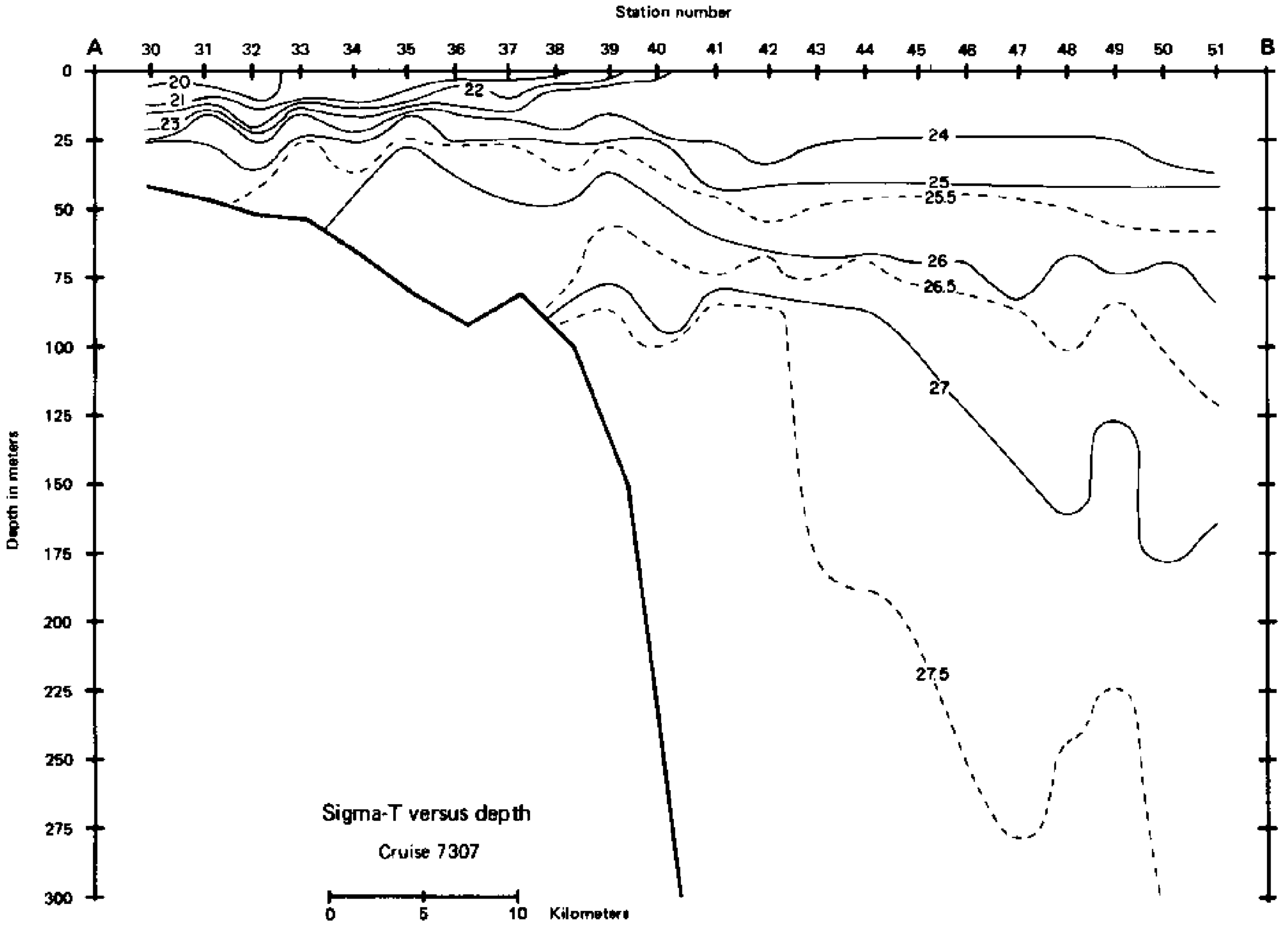


Source: After Bowman and Duedall 1975

Figure 11. Vertical hydrographic sections along line A-B in Map 12, 25-26 July 1973

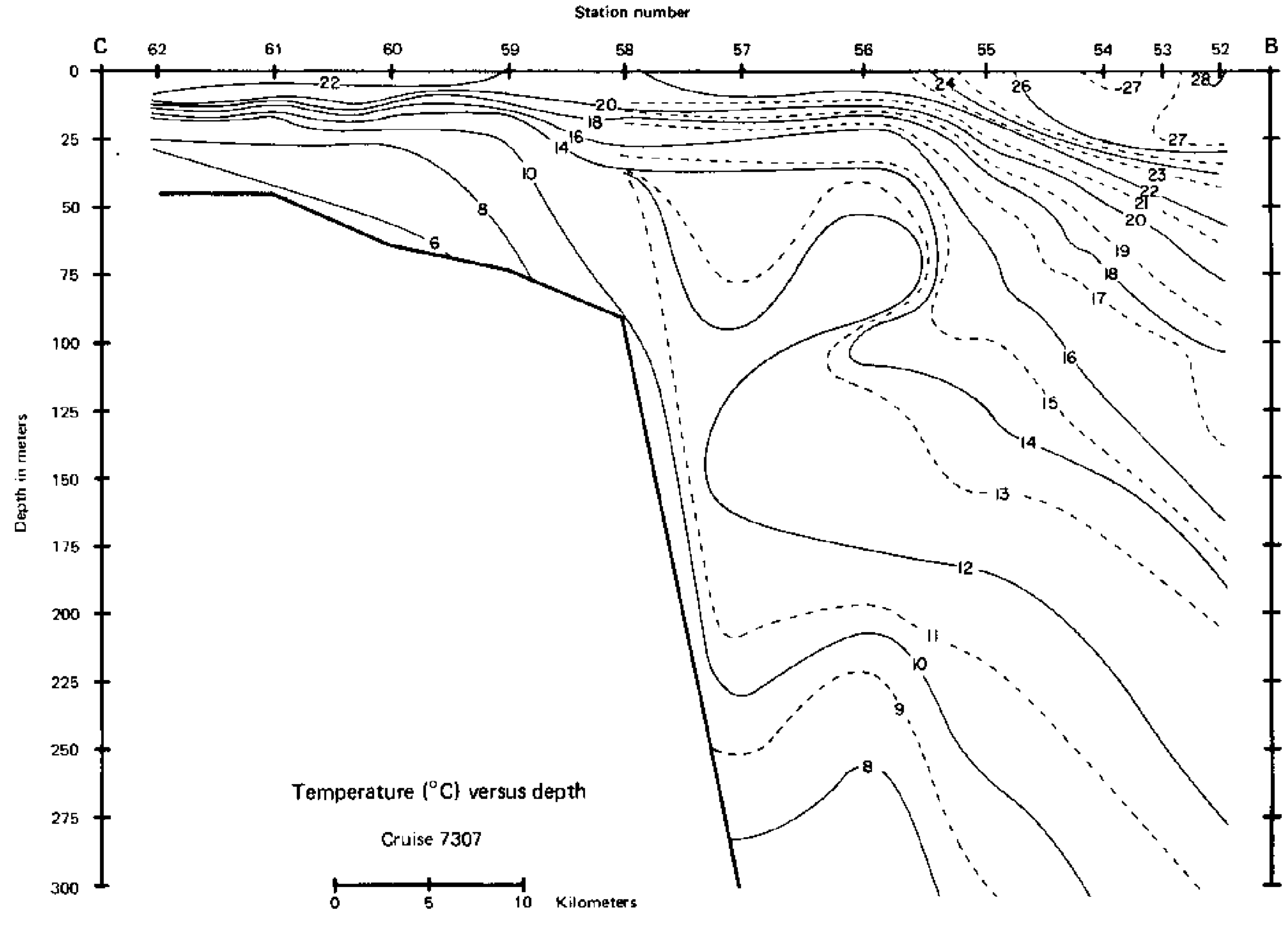




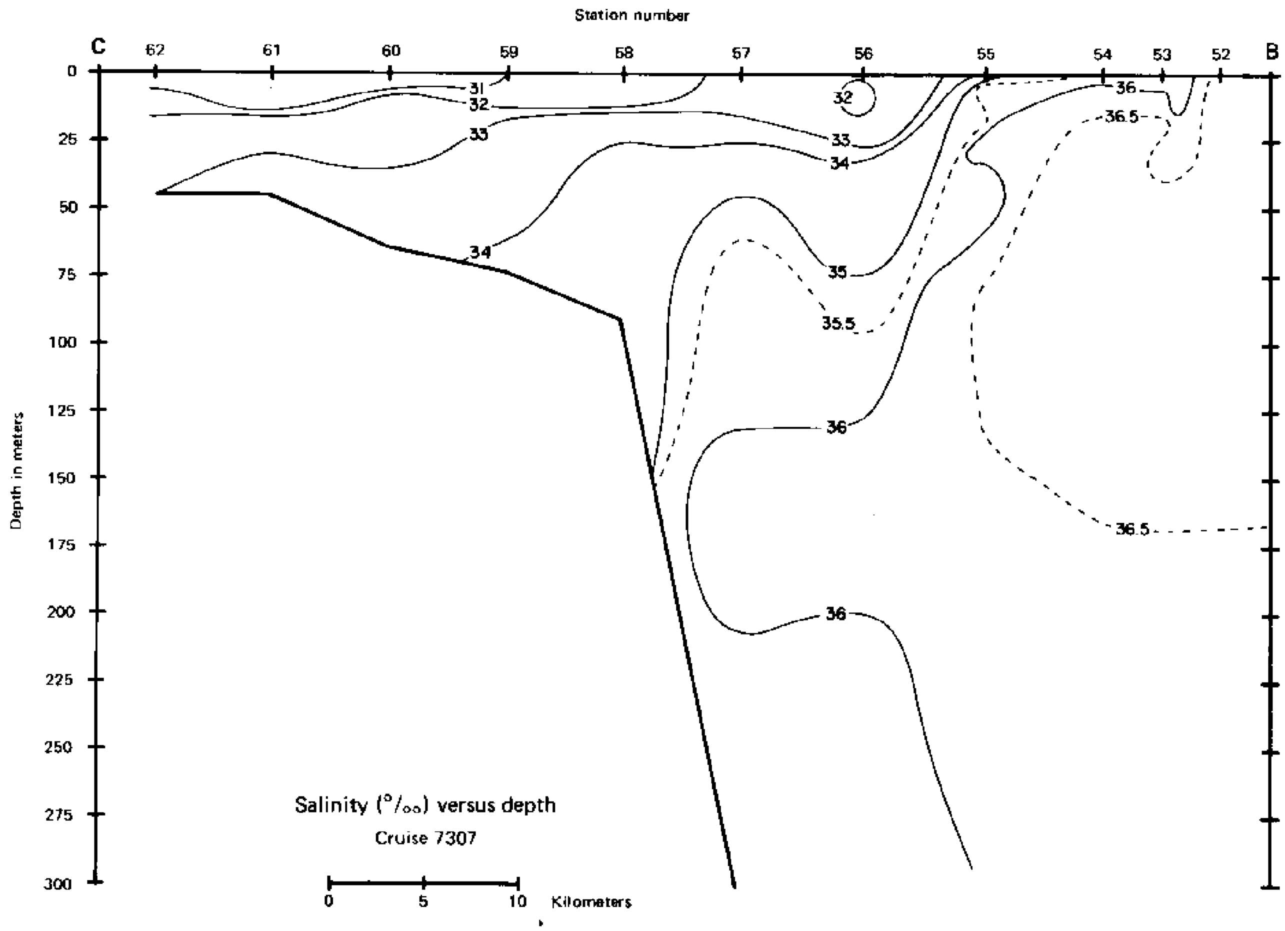


Source: From Bowman and Duedall 1975

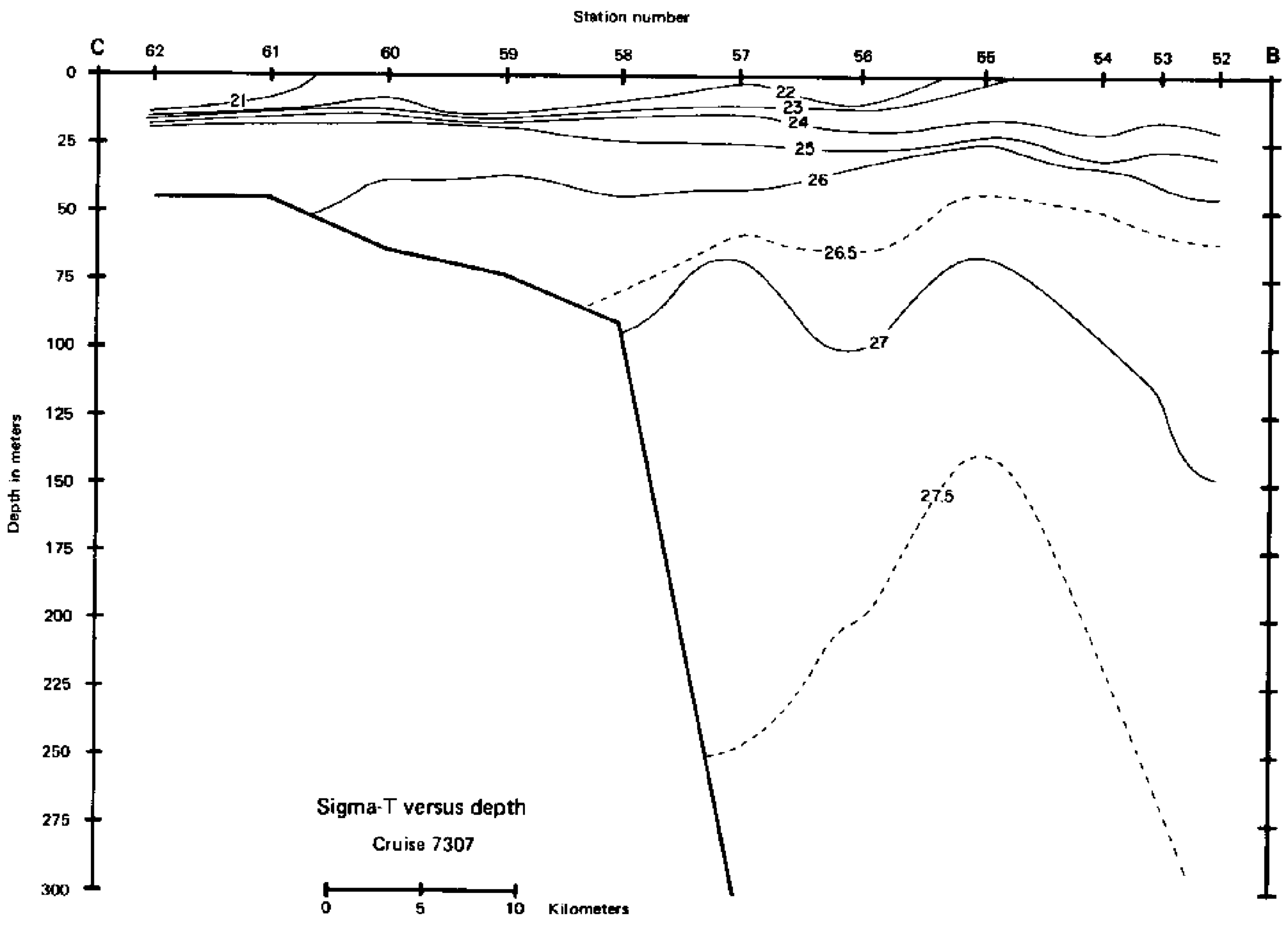
Figure 12. Vertical hydrographic sections along line B-C in Map 12, 26-27 July 1973



Source: From Bowman and Duedall 1975

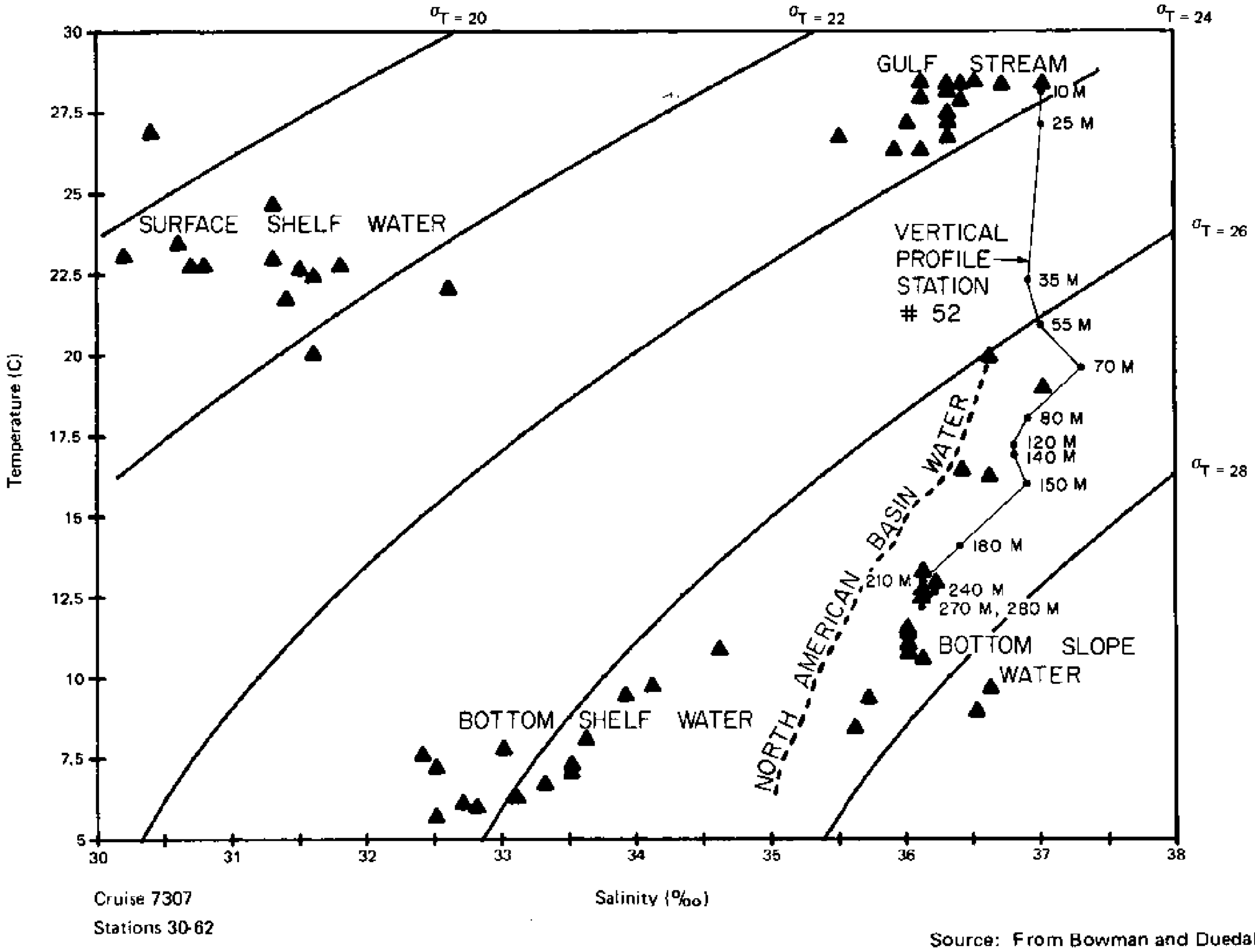


Source: From Bowman and Duedall 1975



Source: From Bowman and Duedall 1975

Figure 13. Temperature-salinity diagram of Gulf Stream meander into Bight, July 1973



Source: From Bowman and Duedall 1975

Gulf Stream Eddies

Mesoscale anticyclonic eddies, which break off the Gulf Stream, have been observed in increasing numbers over the last few years and represent important mechanisms for transferring heat, momentum, and salt from the Gulf Stream into slope waters. These eddies form when a tight Gulf Stream meander, convex inshore, pinches off and detaches a clockwise rotating vortex containing a mixture of warm saline Gulf Stream and Sargasso Sea water.

Fuglister and Worthington (1951) first reported on the formation of a Gulf Stream eddy. Saunders (1971) provides details of the spawning of an anticyclonic eddy near the outer Bight during autumn and early winter of 1969. The eddy, a warm, saline lens floating in slope water, was 100 by 200 km at the surface and 60 by 120 km at 200 m. The eddy was confined to the top 1,000 m in the water column, and surface circulatory currents were on the order of 30 to 70 cm sec^{-1}. Estimates of energy conversions via heat exchange with the atmosphere and dissipation through vertical mixing yielded approximate maximum eddy lifetimes of six months to one year.

Fuglister (1963) estimated that about five to eight cyclonic and anticyclonic eddies are produced each year. Since January 1971 a considerable number of these Gulf Stream anticyclonic eddies have been tracked by scientists of the Naval Oceanographic Office. Eight passed through New York Bight: January 1971, October 1971, October 1972, January 1973, September 1973, December 1973, March-June 1974, June-July 1974. The apparent increase in frequency of occurrence can be attributed in part to the recent use of very high resolution infrared satellite radiometers.

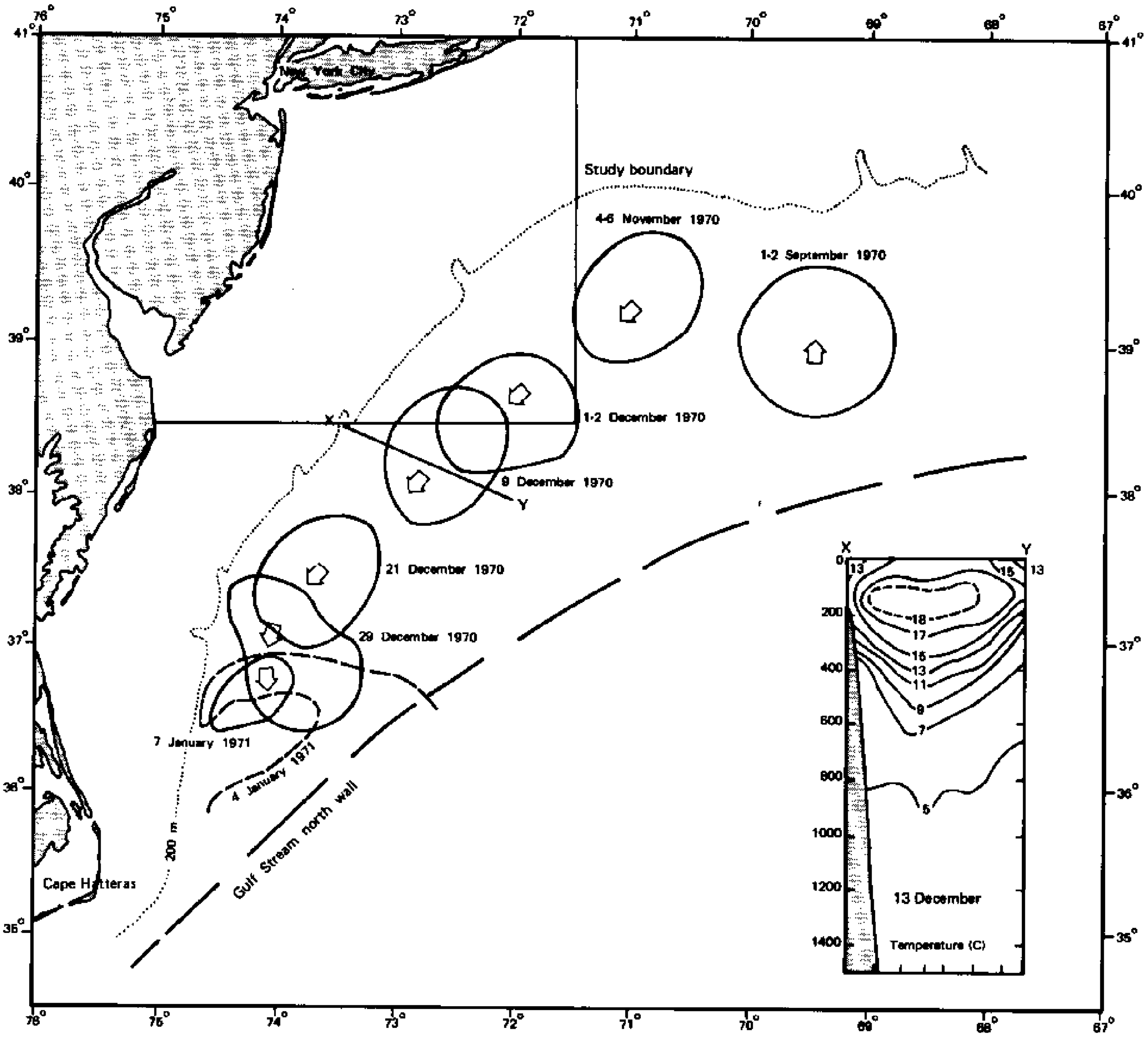
Map 13 illustrates the northwestward and southwestward drift of one eddy that broke off the Gulf Stream in August 1969 and moved through the outer edge of the Bight during December 1970. The insert shows the large horizontal and vertical temperature gradients that persisted for several months inside the eddy.

After formation the anticyclonic eddies drift slowly southward, deflected by the continental slope, at velocities around 5 to 10 cm sec^{-1}, giving them a lifespan of about 150 days. Thompson and Gotthardt (1971) and Gotthardt and Potocsky (1974) first observed the coalescence of eddies with the Gulf Stream near Cape Hatteras during 1970 and in summer 1973. Whether this coalescence was related to the large Gulf Stream meander observed shortly thereafter in the Bight (Bowman and Duedall 1975) is unknown.

The effect of eddies on the biota at the outer shelf edge of the Bight is not understood, although it is believed that an eddy advects its own characteristic ecosystem along with it (Haedrich 1972).

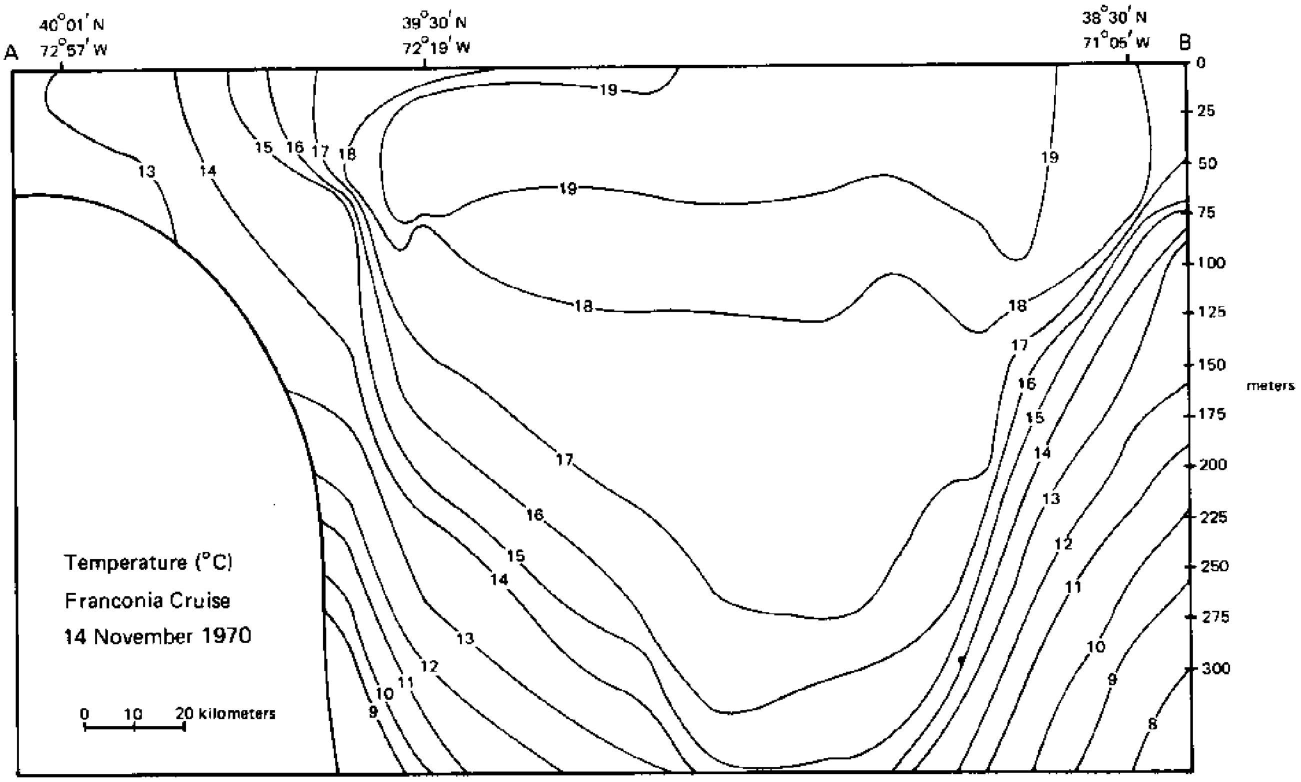
A major disruption in the slope waters of the Bight is shown in the temperature profile in Figure 14. An anticyclonic eddy or the signature of a previous Gulf Stream meander may have been responsible for the presence of warm water and the absence of the slope front in this profile.

Map 13. Drift of anticyclonic eddy



Source: After US Naval Oceanographic Office 1970

Figure 14. Vertical temperature profile along line parallel to A-B in Map 12



Summary

The major characteristics of the hydrography of the Middle Atlantic Bight, including New York Bight, are now well understood. Seasonal cycles of runoff, *insolation,* wind patterns, mean temperature, salinity, and density distributions are well documented. However, important small-scale events such as the exchange of shelf and slope water during all seasons are not nearly so easily observed or quantified. The time and space variability of these phenomena and their relationship to specific wind and storm events need further investigation. Whether the summer cold pool of bottom shelf water is isolated or renewed needs to be resolved.

The influence of mesoscale Gulf Stream eddies in slope waters on shelf water properties is not known. The small-scale dynamics of frontogenesis over the shelf break has not been investigated; little is known about the propagation of low-frequency energy onto the shelf from the deep ocean and its influence on spatial and temporal variability of the frontal zones.

Most of our present knowledge about circulation through the Bight is derived from drift data. Critical is an accurate understanding of the residual circulation in the Bight apex, especially in and below the summer thermocline. Contemporary research is focusing on how specific wind events affect shelf dynamics; much needs to be learned about the details of both longshore and cross-shelf transport mechanisms in unstratified and stratified conditions before definitive conclusions about pollutant dispersal can be reached.

Glossary

Advection: in oceanography, refers to the horizontal or vertical flow of seawater as a current without mixing with the surrounding environment

Convergence zone: region where waters of different origins come together; along a convergence line the denser water from one side sinks under the lighter water from the other side

Coriolis force: apparent force from the earth's rotation, causing moving particles to be deflected to the right of motion in the northern hemisphere, to the left in the southern hemisphere; it is proportional to the speed and latitude of the moving particle and cannot change the speed.

Eddy: circular movement of water, usually formed where currents pass obstructions, between two adjacent currents flowing counter to one another, or along the edge of a permanent current such as the Gulf Stream

Ekman flux, layer, spiral: theoretical representation of the effect of a wind blowing steadily over an ocean of unlimited depth and extent and of uniform viscosity—that is, the surface layer would drift at a 45° angle to the right of the wind direction in the northern hemisphere. Water at successive depths would drift more and more to the right until at some depth it would move opposite to the wind direction; velocity decreases with depth throughout the spiral. This direction reversal occurs around 100 m (depth of the Ekman layer) and the net water transport (Ekman flux) is 90° to the right of the wind direction in the northern hemisphere. A bottom Ekman layer is established by the drag of a current flowing over the ocean bottom.

Envelope: ocean region containing all observed meanders of a current system

Estuarine circulation: two-layer, nontidal flow established by the horizontal density gradient (arising from the intrusion of seawater) in an estuary. A surface seaward flow of brackish water is matched by a landward flow of denser water at depth; intensities of the two flows may be 10 to 50 times the strength of the river flow into the estuary head.

Expendable bathythermograph (XBT): device for measuring temperature versus depth from the sea surface down to 500 m; especially designed for use while a ship is under way. Falling through the water at a fixed rate, a thermistor temperature probe transmits a signal up a fine two-conductor wire attached to spools on the ship's deck. The temperature depth trace is displayed in real time on the ship's recorder. When the wire is entirely payed out, it breaks, dropping the probe to the sea bottom. A modification is the aircraft XBT (AXBT), which uses a radio link from a surface float above the probe to the aircraft overhead.

Flux: rate of transport of fluid properties such as momentum, mass, heat, or suspended matter by eddies in turbulent motion; rate of turbulent exchange

Front: interface or transition zone between two water masses of different density (and, correspondingly, of different temperature and salinity)

Frontogenesis: processes leading to the creation of fronts in the ocean

Geostrophic current: defined by assuming that an exact balance exists between the horizontal pressure gradient and the Coriolis force

Gradient current: defined by assuming that the horizontal pressure gradient in the sea is balanced by the sum of the Coriolis force and the bottom frictional force. At some distance from the ocean bottom the effect of friction becomes negligible; above this the gradient current and *geostrophic current* are equivalent.

Halocline: well-defined vertical gradient of salinity, usually positive (that is, salinity increases downward) on the continental shelf

Horizontal density gradient: rate of change of density with horizontal distance

Hydraulic flow: through a channel, gravity flow that results from a difference between water levels at the channel ends because of a difference in the tide phase or range

Hydraulic gradient: in open channel flow, the slope of the water surface taken parallel to the flow

Insolation (contracted from incoming solar radiation): in general, solar radiation received at the earth's surface

Isohaline: on a chart, line connecting all points of equal salinity; an isopleth of salinity

Isoline (isopleth): line of equal or constant value of a given quantity with respect to either space or time

Isopycnal: on a chart, line connecting all points of equal density; an isopleth of density

Isotherm: on a chart, line connecting all points of equal temperature; an isopleth of temperature

Marsden square: on a Mercator map projection, world between 90°N and 80°S is divided into 10° squares, which are numbered to indicate position based on a system introduced by Marsden in early 19th century for showing distribution of meteorological data on a chart over oceans; each square may be further divided into quarters or numbered 1° subsquares.

Meander: wavelike deviation of a current system boundary from its mean position

Mechanical bathythermograph (MBT): device for obtaining a record of temperature against depth (strickly speaking, pressure) in the ocean, from a ship under way. Its thermal element is a xylene-filled copper coil that actuates a stylus through a Bourdon tube; the pressure element is a copper aneriod capsule that moves a coated glass slide at right angles to the stylus' motion. A double analog record is thus obtained as the MBT is lowered and recovered.

Mesoscale: oceanic features with a size scale of 10 to 100 km

Momentum budget: mathematical equation relating various momentum fluxes within a defined region (for example, across the air-sea interface)

Plume: well-defined intrusion of brackish water from an estuary mouth into a coastal sea

Pycnocline: vertical gradient of density, determined by the presence of a thermocline or *halocline*

Relaxation: return of a system to its original state after an intial perturbation

Residual drift: motion of a water parcel after tidal motions have been subtracted out

Sigma-T (σ_T): conveniently abbreviated value of the density of a seawater sample with temperature T and salinity S

$$\sigma_T = (\rho\text{-}1) \times 10^3$$

where ρ is the value of the seawater density in cgs units at standard atmospheric pressure; for example, if $\rho = 1.02648$, then $\sigma_T = 26.48$

Specific volume: volume per unit mass of a substance, or the reciprocal of density; in oceanography, specific volume is taken as the reciprocal of specific gravity

Steric anomaly (specific volume anomaly): excess of the actual *specific volume* of seawater at any point in the ocean over the specific volume of seawater of salinity 35°/₀₀ and temperature 0°C at the same pressure

Thermocline: in some layer of a water body, a vertical negative temperature gradient appreciably greater than the gradients above and below it; a layer in which such a gradient occurs. Principal thermoclines in the ocean are either seasonal, due to heating of the surface water in summer, or permanent.

Tidal excursion: horizonal distance a parcel of water is moved in a tidal flow between the end of ebb and the end of flood tide

Tidal flux: rate of tidal flow in $m^3\ sec^{-1}$

Tidal prism: difference between the mean high water volume and the mean low water volume of an estuary

Velocity shear: rate of change of velocity with horizontal or vertical distance

Water year: runs from October of the previous calendar year through September of the current year.

Source: Baker, B.B., Jr., Deebel, W.R., and Geisenderfer, R.D., eds. 1966. *Glossary of oceanographic terms.* 2nd ed. Washington, DC: US Naval Oceanogr. Off.

CONVERSIONS

°C x 9/5 + 32 = °F

range °C x 9/5 = range °F

$cm\ sec^{-1} = 0.394\ in\ sec^{-1}$

m = 3.28 ft

km = 0.621 mi

$km^2 = 0.386\ mi^2$

$km^3 = 0.24\ mi^3$

$m^3 = 1.31\ yd^3$

g = 0.035 oz

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