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OCEANOGRAPHY OF THE NORTHWEST ATLANTIC CONTINENTAL SHELF (1,W)

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1. Introduction

This review (Region 1, W) covers coastal and shelf waters of the North Atlantic Ocean from Cabot Strait, situated between Nova Scotia and Newfoundland, to Cape Hatteras, North Carolina (from approximately 47°N to 35°N latitude; Fig. 5.1). It includes the shelf regions off the mouth of the Gulf of St. Lawrence, the Nova Scotian Shelf, Georges Bank and the Gulf of Maine, the southern New England Shelf, and the Middle Atlantic Bight.

The continental shelves throughout the region are broad, especially off Nova Scotia and the Gulf of Maine, and extend more than 200 km offshore. They are cut by a number of deep channels, most notably the Laurentian Channel (>300 m) which runs through the Gulf of St. Lawrence, and the Northeast Channel (>250 m) at the mouth of the Gulf of Maine (Fig. 5.1); each provides an important connection between shelf waters and the open ocean. Several major rivers – especially the St. Lawrence River and the Hudson River – and many more smaller rivers and streams collectively contribute significant volumes of freshwater to the coasts. An

additional volume of freshwater, approximately equal to that from local rivers and streams, is delivered to the shelf regions as part of the coastal limb of the Labrador Current.

Generally speaking, coastal and shelf waters throughout the region support extensive and productive fisheries. Their relatively high biological productivity results from a number of interacting features and processes, including cross-isobath fluxes of nutrient-rich deep waters, which occurs year round, and winter convective mixing. Winter mixing annually replenishes surface nutrient concentrations, setting the stage for important winter-spring plankton blooms that often commence in cold water temperatures ($<1.0^{\circ}\text{C}$), facilitating efficient benthic-pelagic coupling. The spring bloom period is followed by strong vertical stratification throughout the warmer months, established by both freshwater additions and vernal warming of surface layers. Vertical mixing by tides throughout much of the region, amplified by local resonant effects, further stimulate nutrient fluxes that promote high levels of plankton production. In addition, estuarine systems such as the St. Lawrence and Hudson River estuaries, and the Delaware, Narragansett and Chesapeake Bays are both highly productive and significantly impacted by human activities, as are numerous shallow embayments encompassing a wide range of spatial scales.

Ecosystems throughout the northwest Atlantic shelf have been and continue to be influenced by climatic cycles and anthropogenic impacts such as coastal eutrophication and over fishing; these plus present and future oil and gas exploration activities pose significant but poorly understood stresses. Each of these influences creates a number of future scientific challenges, and assessment of new research directions will more than likely be set by pragmatic concerns. Identification of new avenues of marine environmental research, some of which are suggested in this review, have the advantage of being built upon a solid foundation of background knowledge, for this region of the world ocean can be argued to be the best studied of any. Such is the backdrop for our review of this extensive region. We caution that we can offer here only highlights of important oceanographic features, and more often than not, we have been forced to omit altogether discussions of interesting aspects of smaller subregions. Nonetheless, we have found that reviewing the northwest Atlantic shelf region as a highly coupled and interactive system has been both interesting and informative. Our approach in this review thus begins as a general overview of the large-scale physical setting, which we follow with more focused discussions of interdisciplinary aspects of the more-easily identified shelf systems from north to south.

2. Physical Setting

2.1. Physiography

The physical geography of this ocean margin has been reviewed extensively by Emery and Uchupi (1972). The orientation of the coastline and primary land form are controlled largely by the parallel Appalachian, New England and Maritime highlands directly to the northwest. These highlands generally lie several hundred km from the shelf-slope break, with greater distances from southwest to northeast. Their proximity to the coast restricts the overall area of watersheds that drain into the shelf areas, except for major rivers (e.g., Susquehanna, Hudson, and St. Lawrence) from the continental interior that breach these higher elevations. Glacial

scouring has left a complex, incised coastline toward the northeast, with rocky headlands separating frequent small, and occasional larger, estuaries (reviewed in Roman et al., 2000). Toward the southwest the coastline transitions to long, sandy shorelines occasionally breached by larger estuarine systems such as the Hudson, Delaware and Chesapeake.

The numerous estuaries throughout the region effectively trap the vast majority of suspended sediments delivered by their own watersheds as a result of local estuarine circulation patterns (e.g., Meade, 1972; Woodruff et al. 2001), although some escape is possible during rare flood events. These systems may also import fine-grained sediments brought in from adjacent shelf waters. The continental shelves are wide, as already mentioned, but they vary with location, being widest in the northeastern sector, starting at about 250 km in the eastern Nova Scotian Shelf and narrowing to about 30 km at Cape Hatteras. The large embayment of the Gulf of Maine, mirrored by a southeasterly bulge in the shelf off of Cape Cod, which is Georges Bank, effectively enlarges the shelf width to about 400 km. The shelf break shoals irregularly from northeast to southwest, and ranges from 150 m to 50 m depth. Shelf bathymetry varies along this northeast-southwest transect, with numerous mud-bottomed basins and sandier, interbasinal, topographic highs important in the northeast section (Nova Scotian Shelf and Gulf of Maine) grading to the relatively featureless, sandy areas southwest of Long Island. Important incisions are cut into the shelf at the Gully (east of Sable Island on the Nova Scotian Shelf), the Laurentian Channel (entering the Gulf of St. Lawrence), the Northeast Channel (entering the Gulf of Maine) and the Hudson Channel (connecting the Hudson River to the Hudson Canyon on the slope). Similar incisions toward the southwest have been largely buried by more intense sedimentation (e.g., Evans et al., 2000). These incisions provide important flow paths that allow deeper, slope-derived waters to advect closer to shore than normal cross-shelf mixing would allow.

Progressive narrowing of the shelf width from about 150 km off New York to only about 30 km at Cape Hatteras, significantly influences cross-isobath flows and exchanges between the shelf and the open ocean and Gulf Stream (Bigelow, 1933; Churchill and Cornillon, 1991; Bignami and Hopkins 2003; and many others). Off the coast of New Jersey, local topographic highs and relict river deltas have been identified as important factors in persistent upwelling processes (Glenn et al., 1996). A series of major estuarine systems are found off southern New England and in the Middle Atlantic Bight, at the Narragansett, Hudson-Raritan, Delaware and Chesapeake Bays, and the Pamlico Sound (via Ocracoke Inlet) all of which open directly into to the coastal ocean. The coastline through this region can best be described as angular, with long, relatively straight sections of shoreline extending away from either side of major estuaries. The apex of the New York Bight, for example, is composed of two sections of coastline oriented to one another at an angle approaching 90 degrees, and does not possess the steep bathymetry of the Gulf of Maine, nor the cusped embayments of the northern South Atlantic Bight.

2.2. Circulation and Water Masses

The shelf waters of the northwest Atlantic are located in a region of abrupt changes in water temperatures with latitude at the confluence of the Gulf Stream flowing north and the Labrador Current flowing south (Fig. 5.2). In addition, mid-latitude cyclones frequently track across North America and converge in this region (Fig. 5.3), which likely has a significant impact on vertical mixing and nutrient fluxes. Water properties in the region are affected by

both these influences, the Gulf Stream and storms, but they are governed more so by oceanographic processes occurring to the north and “upstream”. Details of the those processes are discussed in the accompanying review of the Northern North America and West Greenland waters by A. Clarke (this volume) and only the major features are summarized here as they affect waters southwest of the Gulf of St. Lawrence.

The large scale physical oceanography of the Northwest Atlantic continental shelf region has been reviewed by Loder et al. (1998). The main features are greatly influenced by dynamics of the North Atlantic subpolar and subtropical gyres, which meet off the Grand Banks of Newfoundland. The major current systems are illustrated in Figure 5.4, and include the Labrador Current, the Gulf Stream, and their adjoining Shelf and Slope Water currents, as described by Csanady and Hamilton (1988) and Chapman and Beardsley (1989). A continuous equatorward coastal current system extends throughout the region, from Newfoundland south to the Middle Atlantic Bight, which interacts with Slope Waters between it and the Labrador Current, north of the Grand Banks, and the Gulf Stream farther to the south and southwest.

The equatorward flowing Labrador Current extends from the southern tip of Baffin Island in the Arctic to the southern tip of the Grand Banks, where it meets the westward-flowing Gulf Stream and North Atlantic Current. The Labrador Current is a cold, relatively fresh, buoyancy-driven coastal current that has its origins on the west coast of Greenland; much of its freshwater derives from Greenland glacial melt (Chapman and Beardsley, 1989). It bifurcates at the Davis Strait, between Greenland and Baffin Island, with one branch flowing north into Baffin Bay, and the other branch(s) crossing Davis Strait, where subsequently the West Greenland Current, the Baffin Land Current from Baffin Bay, and Hudson Bay waters emanating from the Hudson Strait, all come together, further intensifying the flow. Significant additions of freshwater enters from Hudson Strait. Thus, intense winter cooling and freshening by Arctic rivers and ice melt contribute to the Labrador Current’s water properties. The relatively broad Current extends across isobaths from the continental shelf over the continental slope and rise, and comprises what is commonly known as Labrador Slope Water, which resides beyond the shelf break as well as over the continental shelf as an inshore branch. It continues to flow south before branching again into two currents, with most of its transport directed along the outer edge of the shallow Grand Banks; a smaller fraction (ca. 10%; Chapman and Beardsley, 1989) flows across the Grand Banks. This cold and fresh admixture of shelf and slope waters continues to the Scotian Shelf, some of which enters as deep water flows and mixes with the Gulf of St. Lawrence, through the Laurentian Channel, and the Gulf of Maine via the Northeast Channel. Significant volumes of freshwater from the St. Lawrence river is added to the shelf flow.

The general flow of shelf waters continues south as Middle Atlantic Bight water to Cape Hatteras where the shelf width becomes constricted, and cross shelf mixing with the Slope Waters and the Gulf Stream become important (Churchill and Berger, 1998). The shelf and slope waters of the Middle Atlantic Bight maintain their relatively low salinity (<34) augmented by various rivers, some draining heavily urbanized areas (Fisher, 1980) such as the Hudson River, and tributaries that drain via the Chesapeake and Delaware Bays (Chapman and Beardsley, 1989; Lozier and Gawarkiewicz, 2001; Scudlark and Church, 1993; Malone et al., 1996; Magnien et al., 1992; Wang et al., 2001). Mean currents in this region are directed to the southwest with an integrated mean flow on the order of 0.03-0.1 m s⁻¹ (Beardsley et al., 1976; Flagg et al., 1998; Mountain, 2003). While the mean velocity remains about the same throughout

the region, the cross sectional area of the shelf first increases from Nantucket Shoals to Maryland, and then decreases dramatically toward Cape Hatteras. There are also changes in volume transport, from ~ 0.4 Sv at Southern New England to ~ 0.2 Sv off Maryland (Beardsley and Boicourt, 1981; Biscaye et al., 1994), and water mass characteristics that suggest considerable cross frontal exchange at the shelf break (Gordon and Aikman, 1981; Flagg et al., 1998; Gawarkiewicz et al., 2001). As Flagg et al. (1998) state, effectively all of the water entering the Middle Atlantic Bight, through inflows from the north, river discharge at the coast, and cross frontal exchange at the seaward edge, must leave the system by the time the flow reaches Cape Hatteras. Although there is evidence of “leakage” of Middle Atlantic Bight Water into the South Atlantic Bight (Pietrafesa et al., 1994; Grothues and Cowen, 1999), a consequence is that exchange, in both directions, with the slope sea is fundamentally important to the oceanography of the region.

As surface water temperatures increase in the spring along the southern, offshore edges of the Nova Scotian Shelf, Georges Bank and the Middle Atlantic Bight, a seasonal pycnocline develops that isolates a relatively cold water mass – the cold pool – below the seasonal thermocline. First described by Bigelow in 1933, this feature has been the subject of significant attention over the years, and has been reviewed by Houghton et al. (1982), Flagg et al. (1998) and Bignami and Hopkins (2003). The cold pool is an outer shelf phenomenon, largely because tide and wind related vertical mixing can result in warming of deeper waters in the shoaler regions. The foot of the shelf-slope front isolates the cold pool from warmer, more saline, denser slope water. Waters that constitute the cold pool advect along the shelf to the southwest with the mean flow, which results in some of the coolest temperatures in the southern Middle Atlantic Bight occurring in the summer rather than in winter. Analogously, the cold pool advection to the southwest brings additional nutrients to the southern Middle Atlantic Bight from the northern Middle Atlantic Bight. A second result of this advection is that cold pool water experiences a narrowing of the shelf as it approaches Cape Hatteras. Flagg et al. (1998) identified a train of anticyclonic eddies, originating from the cold pool, moving south beyond the shelf break of the southern Middle Atlantic Bight. These eddies represent a significant loss of water from the system – arguably equal to the mean transport of ~ 0.2 Sv calculated for the region near 38° N. Furthermore, Flagg et al. (1998) argue that if cold pool water accumulated the oxidized products of shelf production (Falkowski et al., 1988), these events could result in the export of organic carbon from the shelf for burial in the deep ocean.

Although it is clear that much of the general southward flow in the Middle Atlantic Bight exits the shelf at the confining topography around Cape Hatteras, there is evidence of significant transport – a minimum of 10% of the mean annual transport over the MAB shelf – of Middle Atlantic Bight Water to the South Atlantic Bight (Pietrafesa et al., 1994; Grothues and Cowen, 1999; and many others). Using seven years of data from several stations in the Middle and South Atlantic Bights, Pietrafesa et al. (1994) found that salinity at Diamond Shoals was highly correlated with alongshore wind stress suggesting that wind driven advection of the front between Virginian Coastal Water and Carolina Coastal Water across Diamond Shoals occurs. Pietrafesa et al. (1994) also found Virginia Coastal Water as far south as Frying Pan Shoal – more than 200 km south of Cape Hatteras – which may represent a significant source of freshwater to the northern South Atlantic Bight. This does not seem to be a rare event; Pietrafesa et al. claim that Virginian Coastal Water can be found the South Atlantic Bight during a seven

year record more than 50% of the time.

2.3. Influence of The North Atlantic Oscillation

Labrador Current and Slope Water can be traced as far southwest as the New York Bight; its transport so far to the southwest depends on variable wind and baroclinic forcing, controlled at least in part by large scale alterations in atmospheric circulation patterns produced by the North Atlantic Oscillation (NAO) (Pershing et al., 2001; Drinkwater et al., 2002; Greene and Pershing, 2003). At various times and in varying amounts, under NAO influence, Labrador Slope Water flows southwest between the shelf break and Warm Slope Water, which is of North Atlantic Central Water origin and resides adjacent to the north wall of the Gulf Stream, mixing with Warm Slope Water along the way. Gatién (1976) described Labrador Slope Water as having temperatures and salinities of 4-8°C and 34.3-35, versus Warm Slope Water which has T-S characteristics of 8-12°C and 34.7-35.5. Each water mass reflects the relative importance of Labrador Current water and North Atlantic Central Water. These slope waters adjacent to the shelf break also intrude inshore onto the Scotian Shelf and into the Gulf of Maine as bottom water inflows through deep channels that cut into the shelves (Laurentian Channel, the Gully, and the Northeast Channel; Fig. 5.1). Thus, the deep and bottom water properties of those influxes to inshore shelf regions can vary widely, from warm and salty Warm Slope Water, to cold and relatively fresh Labrador Slope Water. Recent studies, discussed below, have shown that the impacts of these water masses can have important biological implications.

The North Atlantic Oscillation (NAO) has been receiving increased attention in recent years as an important modulator of water mass properties in the Northwest Atlantic Shelf region, and has been implicated in exerting controls on the biology of those waters (Green and Pershing, 2003; Thomas et al., 2003). NAO is a decadal-scale oscillation of wintertime surface atmospheric pressure over the Arctic (Icelandic Low) and the subtropical Atlantic (the Bermuda-Azores High). The long-term mean winter pressures (December to February) are approximately 1000 mb over Iceland and approximately 1021mb over the Azores, differing on average by about 21mb. The NAO Index is a measure of departures from this long term mean difference and as such it characterizes the wintertime westerlies over the northern North Atlantic Ocean (Rogers, 1984). Long-term records back to 1860 show the NAO Index to be highly variable among years, with 5-year running averages exhibiting decadal patterns. These fluctuating NAO patterns force an oceanographic response that holds important consequences for the shelf waters in the northwest Atlantic.

In Low NAO Index years, when north-south pressure differences are least, the north wall of the Gulf Stream is displaced farther south, and concurrently, the southward transport of the Labrador Current is intensified (Loder et al., 1998). At such times, Labrador Slope Water can be traced as far south as the New York Bight, versus periods of NAO Highs when Labrador Slope Water extends only as far south as the Laurentian Channel. In fact, the infamous 1882 tilefish kill off New Jersey has been attributed to low lethal temperatures characteristic of the cold Labrador Slope Water, which was associated with a Low NAO Index (Marsh et al., 1999). Also, it is thought that the anomalously low water temperatures in the Northwest Atlantic shelf waters throughout much of the 1960s resulted from a protracted NAO Low which increased transport of cold Labrador Slope Water to the south (Pershing et al., 2001).

The degree to which the North Atlantic Oscillation can be correlated with the hydrography throughout this shelf region can be quite dramatic. Figure 5.5 illustrates the apparent teleconnections between NAO Low periods and the resultant penetration of Labrador Slope Water to the south, as reflected in bottom water temperatures in the Gulf of Maine, and surface water temperatures in nearshore areas as isolated from one another as the New York Bight and Boothbay Harbor, Maine, USA. The arrival of colder waters in the vicinity of the Gulf of Maine lags behind NAO Low periods by as much as 18 months (Pershing et al., 2001; Drinkwater et al., 2002).

Recently, details of how the 1996 NAO Low influenced shelf waters on the Scotian Shelf and in the Gulf of Maine has been described by Pershing et al. (2001), Drinkwater et al. (2002) and Greene and Pershing (2003). They observed a bottom-water influx along the edge of the continental shelf of colder Labrador Slope Water in the fall and winter of 1997-1998 that displaced Warm Slope Water. By March of 1998 Labrador Slope Water had extended as far as the southwest edge of Georges Bank (ca. 69°W), and had penetrated the Northeast Channel, entering the deep basins in the Gulf of Maine. This hydrographic phenomenon lagged the 1996 NAO Low by 12-18 months. The cold and relatively fresh Labrador Slope Water was detected in deep water samples of the Northeast Channel and interior Gulf of Maine (150-200 m) in January of 1998; it continued to spread throughout the interior of the Gulf of Maine the following summer (Greene and Pershing, 2003).

The Labrador Slope Water penetration to the southwest following the 1996 NAO Low was not only a deep-water phenomenon: its cold surface water signature is also clearly visible in satellite images of sea surface temperature during 1998 and 1999 (Fig. 5.6). Those images show how the positions of Warm Slope Water associated with the north wall of the Gulf Stream, and the southwestward penetration of cold Labrador Slope Water differed markedly between spring of 1998, a year and a half following the short-term 1996 NAO Low Index, and spring of 1999, which followed a High Index year (Fig. 5.6). In 1998, the surface water signature of cold slope water could be seen sandwiched between the shelf edge, indicated by the 200 m depth contour in Figure 5.6, and the much warmer waters associated with the north wall of the Gulf Stream. The cold surface water mass extended southwest well beyond Georges Bank in 1998. The following year, cold Labrador Current and Slope Waters extended only as far southwestward as Sable Island on the Nova Scotian Shelf, and instead, much warmer surface waters abutted the shelf edge.

In addition to altering water temperatures, there is evidence to suggest that fluctuations in the relative amounts of Labrador Slope Water versus Warm Slope Water that intrude onto the continental shelves can be important to nutrient fluxes. For example, Drinkwater et al. (2002) showed close coherence between deep water temperatures and nitrate concentrations on the central Nova Scotian Shelf; lower temperature episodes, of Labrador Slope Water origin, correspond with lower nitrate concentrations (Fig. 5.7). This suggests a significantly lower nitrate concentration in the source waters, Labrador Current Water and Labrador Slope Water. Deep and bottom waters of the Labrador Sea, the source of nutrients in Labrador Current and Slope Water, are known to be low in inorganic nutrients; samples collected along a transect off the coast of Labrador by Clarke and Coote (1988; transect A in Fig. 5.4) reveal that deep Labrador Slope Waters carry approximately 17 μM NO_3 (Fig. 5.8). Much earlier, McGill and Corwin (1964) presented similar results based on a single transect between South Wolf Island,

Newfoundland, and Cape Farewell, Greenland, where maximum nitrate concentrations were between 15 and 16 μM . On the other hand, North Atlantic Central water, which comprises Warm Slope Waters is much richer in nitrate, in excess of 24 μM NO_3 (Fig. 5.9; transect B in Fig. 5.4). See also Figure 5.10, which shows Warm Slope Water nitrate concentrations of about 23 μM in the Northeast Channel. Highest nitrate concentrations over the continental slope are manifest as a mid-depth maximum (Figs. 5.8 and 5.9). Alternations in the magnitude of Warm Slope Water and Labrador Slope Water would therefore be expected to exert some effect on the biological oceanography of shelves onto which those waters intrude. For example, upon arrival at the Northeast Channel and prior to presumed entry into the Gulf of Maine, Labrador Slope Water has anomalously low concentrations of inorganic nutrients as compared with Warm Slope Water (Thomas et al., 2003) which one might expect to reduce the biological productivity of Gulf of Maine waters. A station sampled in 1998 and 1999 in the Northeast Channel as part of the U.S. Georges Bank Globec Program, showed large differences in water properties and in the concentrations of nitrate between 1998 and 1999 (Fig. 5.10; Thomas et al., 2003). In 1998 cold, relatively fresh, and nitrate-poor Labrador Slope Water was penetrating the channel, while in 1999 nitrate-rich Warm Slope Water, of Atlantic Central Water origin was dominating the influx. Satellite-derived estimates of surface concentrations of phytoplankton chlorophyll do indeed suggest that phytoplankton biomass in the Gulf of Maine was lower in 1998 than in either the year before or the year after (Thomas et al., 2003; Fig. 5.11), but sorting out the effects of differing water masses, as perhaps influenced by NAO events, is not entirely straightforward. For example, Jossie and Goulet (1993) and Conversi et al. (2001), using Continuous Plankton Recorder results, reported lower abundances of the copepod *Calanus finmarchicus* in the Gulf of Maine region during the NAO-low years (e.g., the “cold 1960s”), but Greene and Pershing (2000) interpret that phenomenon as an effect of reduced advection of offshore *Calanus* populations to the shelf regions, rather than reflecting a trophodynamic linkage with reduced primary production during NAO lows. On the other hand, Kane (2003), in a study of the abundance and distribution of the copepod *Metridia lucens* in the northwest Atlantic shelf region, was unable to find any correlations with NAO.

2.4. Frontal Features

Shelf and slope waters mix not only by way of deep-water influxes but by shelf-slope frontal interactions as well. Frontal systems dominate throughout the northwest Atlantic shelf region, producing strong spatial and temporal heterogeneity in both biological and physical parameters, and thus producing identifiable biogeographic regions that can be clearly delimited by fronts between hydrographically dissimilar waters. They span a wide range of spatial scales, from tidal mixing fronts (Loder and Greenberg, 1986) to the Gulf Stream front, but perhaps the most prominent of these frontal features is the shelf break front, which separates relatively cool and fresh shelf water from warmer, more saline slope waters. The shelf break front is present throughout the year but its structure varies over an annual cycle (Mooers et al., 1978; Pickart et al., 1999; Frantantoni et al., 2000; Lozier and Gawarkiewicz, 2001), along with its equatorward currents, which vary between 0.2 and 0.5 m s^{-1} (Gawarkiewicz et al., 1996; Linder and Gawarkiewicz, 1998). It is this variability that makes exchange across the shelf break front and biological productivity along it so important to the ecosystem here.

Linder and Gawarkiewicz (1998) studied a climatological cross section of the shelf break front. They found that in winter the front extends from the surface to the bottom in the vicinity of the shelf break (at about the 100 m isobath), thus reducing exchange between shelf and slope waters. In spring, they found that the increased vertical stability afforded by sloping isopycnals at the frontal zone region provides a region of increased light availability (shown also by Marra et al., 1982 and Malone et al., 1983), allowing a relatively early spring bloom compared to adjacent waters. In summer, stratification removes the sea surface temperature signature of the front but not the salinity structure. Phytoplankton production at the shelf break increases in summer as result of shoaling of the frontal zone and concomitant flux of nutrients from deeper nutrient-rich waters into the surface layers (Marra et al., 1990). Ryan et al. (1999) showed that in spring the enhanced chlorophyll concentrations evident in ocean color satellite imagery at the shelf break front typically develop in nutrient depleted water associated with meanders of the front along the 100 m isobath. These meanders propagate along the shelf break front at about 9 km day^{-1} , producing steeper isopycnal slopes at the troughs (shoreward intrusions) and are associated with seaward advection of shelf water and newly upwelled water from about 20 m depth along isopycnal surfaces, which is responsible for the enhanced chlorophyll concentrations. In summer the front is a subsurface feature, with the seasonal pycnocline isolated from surface waters, and equatorward current velocities exceeding those in winter. The seasonal transformation of the isopycnal structure is thought to contribute to a subsurface salinity maximum noted by several researchers (e.g., Gordon and Aikman, 1981; Churchill, 1985; and others).

Lozier and Gawarkiewicz's (2001) analysis of drifters released on Georges Bank shows that nearly all were entrained in the shelf break frontal jet current and transported to the south. Often, the drifters were detrained/re-entrained numerous times along the front suggesting that exchange across the front is possible along its entire length. However, the general sense from this work is that transport of drifters is to the southwest and offshore; few drifters (only about 10%) were transported shoreward any considerable distance and those that did, did so during summer.

On the seaward side of Georges Bank, interaction between Gulf Stream warm-core rings and the shelf break front influences both water mass properties and chlorophyll distributions (Ryan et al. 2001). Entrainment of shelf water and elevated chlorophyll concentrations offshore into the ring can result in shelf water characteristics completely encircling the ring. In addition to the actual entrainment, Ryan et al. (2001) noted enhanced chlorophyll along the 60 and 100 m isobaths south of Georges Bank, the location of maximum seaward extent of shelf water characteristics, associated with divergent cross-shelf flow and sea surface temperature minima, consistent with upwelling.

The location and seasonal variability in sea surface temperature fronts calculated from 12 years of satellite data shows strong seasonal and latitudinal differences in the northwest Atlantic shelf region (Ullman and Cornillon, 1999). The shelf-break front is observed in the vicinity of the 200 m isobath from Cape Hatteras to the Scotia Shelf in all seasons except summer, when surface heating removes its signature. In summer, strongest and most persistent surface thermal fronts are present in the Gulf of Maine along the seaward edge of the Eastern Maine Coastal Current, along the edge of the Scotia Shelf and around Nantucket Shoals and Georges Bank where tidal mixing over the shallow bathymetry separates well mixed water from stratified

regions over deeper water. Most (82%) of the non-seasonal variability in SST in the region of the Eastern Maine Coastal Current occurs on time scales of 1-2 months, with little occurring at shorter time scales, suggesting a link with low-frequency variability in the geostrophic flow over Jordan Basin (Bisagni et al., 1996). In winter, surface temperature fronts present on the inner and middle shelf throughout the study area separate nearshore cooler surface water from offshore warmer regions. These likely result from seasonal cooling over shallow regions and result in a reversal of the near-shore thermal gradient over seasonal cycles (Thomas et al. 2002). Between Cape Hatteras and Long Island, surface thermal fronts are weaker and seasonally dominated by those along the inner shelf and shelf break in winter (Ullman and Cornillon, 1999). Summer surface heating eliminates the SST signature of most shelf fronts in this southern portion of our study region, but not the salinity signature.

2.5. The Gulf Stream

The Gulf Stream is the most dominant feature in the northwestern Atlantic Ocean, and influences the dynamics of the adjacent continental shelf waters. Its location offshore ranges from as close as 30 km to the shoreline off Cape Hatteras, to much more widely varying distances offshore as it flows northeastward (Fig. 5.2). The Gulf Stream in the South Atlantic Bight (south of Cape Hatteras) can exist in one of two modal states: the weakly deflected state, with small meander amplitude and small warm filaments rarely penetrating beyond mid-shelf in the South Atlantic Bight and with the Gulf Stream inshore of the 600 m isobath between Charleston and Cape Hatteras; or the strongly deflected state, with the position of the Gulf Stream seaward of the 600 m isobath, large meander amplitude and warm filaments often penetrating to the inner-shelf (Bane and Dewar, 1988). Near Cape Hatteras, the Gulf Stream is aligned with the 100 m isobath in the weakly deflected state, whereas its position in the strongly deflected state is 15-20 km farther offshore (Glenn and Ebbesmeyer, 1994a).

Frontal eddies, arising in meanders of the Gulf Stream, are biologically productive features and propagate toward Cape Hatteras every 4-5 days. Meander amplitude decays as the Gulf Stream approaches Cape Hatteras (Lee et al., 1991) and it is believed that frontal eddies are destroyed as they encounter this highly sheared region. Glenn and Ebbesmeyer (1994a, 1994b) discussed a field program that tracked a frontal meander and associated cold dome in the South Atlantic Bight. They found that cold domes associated with meanders and frontal eddies can propagate past Cape Hatteras into the Middle Atlantic Bight and might thereby create situations in the Middle Atlantic Bight that promote higher primary productivity which has its origins in the South Atlantic Bight. While the cold dome moves downstream past Hatteras, the associated warm filament, they found, could be stranded on the shelf in the South Atlantic Bight to be eventually re-entrained into the Gulf Stream. Furthermore, Glenn and Ebbesmeyer (1994b) speculated that this process might be more successfully when the Gulf Stream was in its strongly deflected mode.

2.6. Sedimentological Characteristics and Processes

The continental margin in our northwest Atlantic study region contains sedimentary deposits of up to several thousand meters in thickness, which date back to the initial rifting of

this part of the Atlantic Ocean during the Mesozoic. Surficial sediments in the region are heavily influenced by reworking of Holocene glacial debris, with little modern delivery of sediment from land. Sands dominate on the open shelf, with occasional deep areas (e.g., Hudson Canyon, Mud Patch off Martha's Vineyard) containing higher proportions of silts and clays. Deep basins on the Nova Scotian Shelf and in the Gulf of Maine are finer grained, with silts and clayey materials abundant at most depths. The region is intensely trawled and sediment resuspension at depths greater than 100 m may well be controlled largely by commercial fishing (Churchill, 1989).

The bottom sediments play important roles in the functioning of northwest Atlantic continental shelf ecosystems. As the repository for material settled from the water column, they serve as storage and processing sites for reactive and unreactive materials contained in the settling detritus. These sedimentary environments serve roles as habitat for benthic organisms, an important zone for denitrification, and a zone for temporary or long-term storage of organic matter received from the water column.

The distributions of benthic animals generally correspond well with sediment grain size. From classic studies in shallow environments (Sanders, 1956) to more modern approaches using landscape ecology approaches (e.g., Zajac, 2001), it is clear that the proximal variable of substrate type, usually parameterized as grain size, is the dominant controlling variable. This region has little sediment delivery, so that water column energy is the major forcing function controlling bottom sediment grain size. Thus water depth, wave action and bottom currents are highly influential in structuring bottom type. While factors such as food supply, predation, temperature and water pressure are likely also of influence, our understanding of their roles is insufficient to distinguish them from the influence of habitat type.

Shelf sediments have been hypothesized to host the most important sink for oceanic nitrogen-denitrification (Seitzinger and Giblin, 1996), and this process has been studied in both the Gulf of Maine and along the Mid-Atlantic Bight areas. Christensen et al. (1987; 1996) used a combination of direct measurements in the sediments and ratios of water column nutrients to show that denitrification in the Gulf of Maine removes about one fourth of the combined nitrogen delivered into the Gulf from offshore. This removal results from a combination of bioirrigational input of water column nitrate into the sediments and coupled nitrification-denitrification that derives from ammonium regenerated from settled organic detritus. They predicted that the relative importance of these two processes should favor the latter in shallower areas. In agreement with this prediction, Hopkinson et al. (2001) found that most denitrification in muddy sediments at 30-75m depth in the Gulf of Maine was supported by mineralization of detritus, and that it accounted for 60% of the total nitrogen remineralized. Laursen and Seitzinger (2002) obtained similar results at the 15m deep, sandy, LEO-15 site off New Jersey, finding virtually all of the denitrification to result from coupled nitrification-denitrification of remineralized organic detritus. The overall denitrification rates on the shelf area addressed in this paper has been estimated via model calculations to be greater than could be accounted by riverine and atmospheric delivery of nitrogen, and instead requires at least half of the nitrogen to be derived from upwelling of oceanic waters (Seitzinger and Giblin, 1996).

Organic detritus that reaches the bottom may not remain there permanently, and this region has provided abundant evidence of subsequent redistribution of biogenic material. The

U.S. Dept. of Energy sponsored a number of programs examining this possibility in the region from Cape Cod to Cape Hatteras, called the Shelf Edge Exchange Program (SEEP) and Ocean Margin Program (OMP). These studies (reviewed by Walsh et al., 1988; Biscaye et al., 1994 and Verity et al., 2002) were driven by the observation (see Walsh, 1991) that the sandy shelf sediments provided little evidence for organic matter accumulation, while adjacent slope sediments provided a more likely depocenter of higher concentrations. Field and modeling studies provided abundant evidence that some fraction of shelf primary production was subsequently washed off and transported to the adjacent slope, but that this fraction was a very minor fraction of the total primary production on the shelf (Biscaye et al., 1994; Verity et al., 2002). Charette et al. (2001) arrived at similar conclusions for the shelf northeast of Cape Cod, based on ^{234}Th work. This export of organic carbon from the shelf likely derives from resuspension and transport of settled detritus. Churchill et al. (1994) have reported that wave energy sufficient to enable this resuspension occasionally impinges at bottom depths as great as 131m. Sediment trap collections showed that much of this exported material was of diatom origin (Falkowski et al., 1994). Radioisotope studies indicate that this detritus spends an average of several years in shelf sediments (Bacon et al., 1994), and is likely mixed into the sediments by bioturbation before being resuspended and transported offshore. The exported material is thus degraded and aged compared with shelf sediment organic matter. Nevertheless, these export fluxes of digestible material are sufficient to support benthic biota in slope sediments at levels as great as found in productive estuaries (Aller et al., 2002), and as lateral fluxes they exceed the vertical fluxes from the water column overlying the slope sediments (Anderson et al., 1994). The southward movement of water along the shelf, combined with the narrowing of the shelf and entrainment in the seaward-flowing Gulf Stream at Cape Hatteras, leads to intensification of this export and subsequently intense benthic communities and carbon burial on the slope at the southern end of the region covered in this chapter. This evidence indicates that settling and resuspension can have dramatic impacts on benthic processes at distances of at least tens of km from the origin of the primary production.

It is interesting to speculate here that the apparent increase in water column nitrate concentrations at intermediate depths from the Labrador Sea (Fig. 5.8) to the Slope Sea of Georges Bank and the Middle Atlantic Bight (Fig. 5.9) is a reflection of this shelf-slope exchange, which enriches deep and intermediate depth nutrient concentrations with increasing distance downstream and to the southwest.

3. Regional Shelf Systems

In addition to important Slope Water influences on the chemical and biological oceanography of shelf waters, and export of shelf-derived primary production and regenerated nutrients to slope waters, a number of more complex processes interact on the shallow banks, deep channels and gulfs throughout the region to affect coastal nutrient dynamics and both the magnitude and style of biological production. The following sections focus on those major, biologically-productive shelf regions: the Nova Scotian Shelf; Georges Bank; the Gulf of Maine region; the southern New England shelf and New York Bight; and shelf waters south to Cape Hatteras (Fig. 5.1). On the whole these sub-regions, particularly those south and west of the Nova Scotian Shelf, have benefitted from a wealth of scientific studies over the past several

decades, and in many ways, they are among the best studied coastal areas of the world ocean.

3.1. Nova Scotian Shelf

The broad Nova Scotian Shelf is characterized by a number of deep basins and channels. It receives a mixture of waters from the Gulf of St. Lawrence and the Labrador Current, which makes Scotian Shelf Waters relatively fresh and seasonally very cold. The Scotian Shelf also receives deeper continental slope waters from beyond the shelf edge (Loder et al., 1998). Seasonal heating of the surface layers traps a cold intermediate water layer beneath the seasonal thermocline and warmer, but saltier (and hence denser) bottom waters that intrude from beyond the shelf edge, thus creating three distinct water types. This formation of a cold intermediate-depth water layer on the Scotian Shelf forms the origin of the “cold pool” (discussed earlier) that extends along the shelf edge to the Middle Atlantic Bight. Frontal dynamics at the shelf break, localized areas of vigorous tidal mixing and upwelling, such as off southwestern Nova Scotia, and seasonal plankton blooms, all contribute to the region’s nutrient dynamics and biological productivity.

Biological-physical coupling and resulting biological productivity of shelf waters off Nova Scotia are in some ways similar to the Grand Banks of Newfoundland farther to the east. Each system is located between 41° and 44° North, has relatively shallow depths, and is affected by relatively fresh cold-water currents from the north. Tidal mixing processes differ between the two regions, however, in that the Grand Banks has smaller tides than do either the Nova Scotian Shelf or Georges Bank. By comparison, the Grand Banks appear to be highly biologically productive in that they have supported a rich fishery for centuries, but despite the purported importance of fisheries production in this system, there is a relatively scant body of literature on plankton production dynamics. Available information indicates that primary production is actually lower than one might expect. For example, Prasad and Haedrich (1993) estimated primary production on the Grand Banks (based on satellite ocean color measurements with the Coastal Zone Color Scanner) in April and May (during the spring bloom) to be on the order of 1,000 mgC m⁻² d⁻¹, and about 300 mgC m⁻² d⁻¹ the rest of the year; thus, annual production averages about 200 gC m⁻² yr⁻¹. Anderson and Gardner (1986) present vertical cross sections in May on the southern portion of the Banks, which indicate that nutrients are depleted from top to bottom over the shallow Banks (50 m), and that the Bank is vertically well-stratified this time of year (pycnocline at ca. 20 m). Cross-shelf influxes of deep water nutrients appear to be limiting on the Banks, in contrast to the Scotian Shelf which is more tidally energetic. In recent years fish stocks have been severely depleted which has led to drastic management practices, including the complete closure of the Newfoundland cod fishery in the early 1990s.

The Scotian Shelf proper, between the Laurentian Channel and the Northeast Channel, has long been known to be biologically productive, based on fisheries landings and observations of large populations of surface-feeding sea birds (Brown et al., 1975). Fournier et al. (1977) reported that phytoplankton biomass and primary production, measured using ¹⁴C incubations, varied with position along a single transect south of Halifax, N.S., being highest about 90 km offshore, in association with the shelf break front. Annually-averaged primary production along their transect was estimated at 96 gC m⁻² yr⁻¹. This relatively low level of production was the average of four cruises, each of which each missed the spring bloom period, which, they pointed

out, would have significantly augmented their estimates. Their findings of elevated production at the shelf break front confirmed earlier observations and speculations offered by others of enhanced biological production at the shelf break off the east coast of North America, where stocks of zooplankton, benthos and fish are reported higher than in waters on either side (Clarke, 1940; Grice and Hart, 1962; Sanders et al., 1965; Schroeder, 1955). Fournier et al. (1977) showed that about 80% of the nitrogen requirements of the shelf phytoplankton could be met by recycling by zooplankton, and that the remaining nitrogen is likely delivered to the shelf by fluxes of slope waters from beyond the shelf break. This important conclusion lent support to the earlier hypothesis by Riley (1967) of the significance of nutrient fluxes; prior to this time, the nutrient requirements of coastal phytoplankton populations was believed to be land-derived. Fournier et al. (1977) also showed in this brief but important paper that the inner basins of the Scotian Shelf are not directly dependent on a cross-frontal flux of nutrient-rich slope waters, since their deeper waters are supplied with slope water that flows in through deep channels which can be subsequently mixed upward by internal waves and tidal mixing.

In a following study conducted during winter, Fournier et al. (1979) showed that elevated primary production can result episodically at times when there is a decrease in the steepness of the isopycnals at the shelf break front, which reduces the mixed layer depth, thus increasing the light environment in already nutrient-rich winter waters. Such events, they argued, can increase primary production there by about 25%. Fournier and his co-workers went on to show that a similar frontal phenomenon is responsible for enhanced primary production on the southwest Nova Scotian Shelf, in this case as a result of a tidal mixing front (Fournier et al., 1984).

Tidal ranges increase from east to west across the Scotian Shelf and into the Gulf of Maine, with some of the highest tides in the world found at the upper reaches of the Bay of Fundy (Fig. 5.12). Tidal current speeds on the southwest Nova Scotian Shelf and over Browns Bank are on the order of 100 cm s^{-1} and produce significant vertical mixing of the water column (Garrett et al., 1978). Shelf waters off southwest Nova Scotia are conspicuously colder than surrounding areas, especially during the warmer months. Vigorous tidal mixing, combined with the flow of cold Scotian Shelf water from the east (as the Nova Scotia Current), and localized upwelling, maintain cold temperatures here year-round. This tidal mixing in combination with upwelling off southwest Nova Scotia (Lauzier, 1967), a result of centrifugal effects of tidal currents following the curved bathymetry (Garrett and Loucks, 1976) and/or longshore density gradients (Smith, 1983), act together to promote significant nutrient additions onto the shelf (Fournier et al., 1984). As a result elevated phytoplankton biomass is routinely found at the frontal boundary between the vertically mixed coastal waters and stratified offshore waters (Denman and Herman, 1978; Fournier et al., 1984). The case for upwelling off southwest Nova Scotia is further supported by observations of a patch of colder water offshore, distinct from the cool coastal waters adjacent to southwest Nova Scotia, routinely seen in satellite images of sea surface temperature (e.g., Fig. 5.13); this is in keeping with the observation by Fournier et al. (1984) of offshore patches of elevated surface nitrate concentrations. (A similar, consistent cold-water patch is observed in the vicinity of the Great South Channel in the western Gulf of Maine; Fig. 5.13).

The commercial fisheries landings on the Scotian Shelf have declined dramatically in the past 30 years – landings of demersal fish on the eastern portion of the Scotian Shelf declined from a maximum of 450,000 metric tons (MT) in 1973 to less than 15, 000 MT in 1997

(Zwanenburg et al., 2002). This sharp decline prompted a moratorium on cod in 1993, which remains in effect today. Catches of pelagic species fell during the late 1970s and remain at about 50,000 MT tons today. In contrast, landings of demersal fishes on the western Scotian Shelf have been significantly lower than in the east throughout the last 30 years, and only rarely exceeded 100,000 MT; the catch began to drop in the 1990s and is at less than 50,000 MT today. Catches of pelagics, principally herring, in western Scotian Shelf waters, on the other hand, have been consistently higher than demersals over the past 30 years, fluctuating between 150-300,000 MT (Zwanenburg et al., 2002). As the commercial catches on the Scotian Shelf declined, there has been a commensurate switch in species composition (Zwanenburg et al., 2002) similar to that also observed on Georges Bank and in the Gulf of Maine (discussed below).

3.2. The Gulf of Maine

The Gulf of Maine covers a broad (international) area between Cape Cod, Massachusetts, and southwestern Nova Scotia (Fig. 5.1). Georges and Browns Banks effectively isolate the Gulf from the open Northwest Atlantic Ocean, forming a semi-enclosed continental shelf sea. At depths greater than 100 m, the exchange of waters between the Gulf and the North Atlantic is confined to the deep (>300 m) Northeast Channel separating Georges Bank from Browns Bank and the Nova Scotian Shelf. The Northeast Channel connects with the interior of the Gulf by way of three major deep basins: Georges, Jordan, and Wilkinson, which are isolated one from another below the 200 m isobath; Georges Basin, the deepest of the three (370 m), forms an irregular *cul de sac* behind Georges Bank at the end of the Northeast Channel. These physical characteristics of deep basins and limited deep-water exchanges with the open Atlantic are coupled with other important features and processes that act together to control the general oceanography of the Gulf, including nutrient fluxes and biological productivity. These features and processes include: vertical mixing by tides (Garrett et al., 1978); the seasonal cycle of heating and cooling which leads to winter convection and vertical stratification in summer; pressure gradients from density contrasts set up by deep water inflows and lower salinity waters (Brooks, 1985); and influxes of the cold, but fresher waters associated with Scotian Shelf Water (Smith, 1983).

Tidal ranges in the Gulf of Maine are among the highest in the world ocean, and consequently generate swift tidal currents. Tides in the Gulf decrease from northeast to southwest (Fig. 5.12) and the resulting differences in intensity of tidal mixing (e.g., Loder and Greenberg, 1986) exert a strong influence on the spatial pattern of hydrographic structure in the Gulf, nutrient delivery to the euphotic zone, benthic-pelagic coupling (Townsend et al., 1992a), and, ultimately, upon biological productivity. Summer satellite images of sea surface temperatures in the interior Gulf of Maine commonly show a distinct thermal front separating the cold, tidally-mixed surface waters of the eastern Gulf from the warmer, vertically-stratified waters of the west (Fig. 5.13).

Vertical mixing by tides and winter convection, coupled with influxes of fresher waters from rivers and Scotian Shelf Water, and dense Slope Waters that enter the Gulf at depth through the Northeast Channel, create distinct water masses (Hopkins and Garfield, 1979) that are important to the overall oceanography of the Gulf. Like Scotian Shelf waters just discussed, the extent of winter convection in the Gulf of Maine is likely variable among years, depending on

winter weather conditions, but convective mixing is generally limited to a depth corresponding to the top of the dense bottom water, of Slope Water origin (Brown and Beardsley, 1978). Seasonal heating of the surface layers then traps a cold intermediate water layer beneath the seasonal thermocline and above the warmer, but saltier (and hence denser) bottom waters, thus creating three distinct water types, termed Maine Surface Water, Maine Intermediate Water and Maine Bottom Water (Hopkins and Garfield, 1979). These three layers are most obvious over the deeper offshore Gulf waters, away from tidally-mixed coastal waters; they become eroded by tides throughout the summer and fall, disappearing first from the eastern Gulf of Maine.

The mean circulation in the Gulf of Maine-Georges Bank region is generally cyclonic, driven by density contrasts between slope waters residing in the three offshore basins, and fresher waters along the coast that are fed principally by discharges from the St. John, Penobscot, Kennebec/Androscoggin, and Merrimac Rivers (Brooks, 1985; Xue et al., 2000). River discharges account for only about half of the freshwater budget for the Gulf of Maine, however; the remaining half enters the Gulf as a surface flow of relatively cold, low salinity Scotian Shelf Waters (Smith, 1983). Beardsley et al. (1997) and Lynch et al. (1997) described the generalized surface circulation in the Gulf as dominated by a buoyancy-driven coastal current system that flows counter-clockwise around its edges. In addition, topographically rectified tidal currents are important in the Gulf (e.g., Loder, 1980), and contribute to clockwise circulation patterns around Browns and Georges Banks, and Nantucket Shoals south of Cape Cod.

The coastal current system in the Gulf of Maine has been argued to be important to the overall nutrient budget and biological oceanography of the Gulf (Townsend et al., 1987; Brooks and Townsend, 1989; Townsend, 1998). Vertical nutrient fluxes are driven by vigorous tidal mixing in the northeastern Gulf, creating summertime surface nitrate concentrations $>7 \mu\text{M NO}_3$ (Townsend et al., 1987); these waters constitute an important component of the eastern Maine coastal current (EMCC) system. Pettigrew et al. (1998) describe the EMCC as the cold band of tidally-mixed water that originates on the southwest Nova Scotian shelf, crosses the mouth of the Bay of Fundy, and continues along the coast of eastern Maine to the offing of Penobscot Bay. Before it reaches Penobscot Bay, the EMCC is often directed away from the coastline and out over the central Gulf of Maine as a plume-like feature of colder water, which is clearly visible in satellite images of sea surface temperature (e.g., Fig. 5.13). The exact trajectory of the EMCC is variable (Brooks and Townsend, 1989; Bisagni et al., 1996; Lynch et al., 1997; Pettigrew et al., 1998). A portion of the offshore-directed plume may be entrained in the cyclonic gyre over Jordan Basin, with the remaining portion entering an anticyclonic eddy at the distal end, bringing EMCC-plume waters back toward the Maine coast where it continues as part of the western Maine coastal current (Pettigrew et al., 1998).

The EMCC and its offshore plume feature of nutrient-rich water are important to the species composition and abundance of plankton in the offshore waters of the Gulf. Townsend et al. (1987) described the sequence of processes operating in the EMCC-plume as beginning with vigorous tidal mixing in the eastern Gulf and the injection of new nutrients at the upstream end. As those waters flow to the southwest they become increasingly vertically stratified, and nutrient concentrations decrease with increasing distance downstream, concomitant with an increase in phytoplankton biomass and developmental stages of copepods. This pattern of an offshore patch of high phytoplankton biomass at the distal end of the plume was noted as early as Bigelow (1926). The magnitude of planktonic production driven by the EMCC-plume is significant:

Townsend et al. (1987) estimated that approximately 44% of the inorganic nutrient flux (to surface waters) required to meet estimated levels of new primary production for the entire Gulf of Maine can be attributed to the EMCC-plume system. Larval herring spawned in eastern Maine are carried by the EMCC and benefit from its productivity (Townsend, 1992), and the dynamics of this current system may be important in explaining the distributions of episodic coccolithophore blooms (Townsend et al., 1994a). During summer, the offshore, nutrient-rich plume feature is one of two sites in the Gulf of Maine (the mouth of the Bay of Fundy being the other) of consistently high cell densities of the toxic dinoflagellate, *Alexandrium fundyense* (Townsend et al., 2001).

Overall, the Gulf of Maine has a reputation of high biological productivity, with the general nature of the biological oceanography of the Gulf and its offshore banks having been known since the early works of Henry Bryant Bigelow (Bigelow, 1926, 1927; Bigelow et al., 1940). Levels of primary production in offshore waters, the least productive areas in the Gulf of Maine, average about $270 \text{ gC m}^{-2} \text{ yr}^{-1}$ (O'Reilly and Busch, 1984; O'Reilly et al., 1987). The principal source of nutrients that support this offshore primary production is generally thought to be the influx into the Gulf of nutrient-rich deep Slope Water through the Northeast Channel (e.g., Fig. 5.10; Ramp et al., 1985; Schlitz and Cohen, 1984; Townsend, 1991; Townsend, 1998). Once delivered into the Gulf, the high concentrations of inorganic nutrients that accompany these deep slope water intrusions are delivered upward to the surface by various mechanisms and thus eventually are made available for planktonic primary production. The major mechanisms that supply nutrients to surface waters include: vertical mixing by tides and upwelling in the eastern Gulf and on the southwest Nova Scotian Shelf; fluxes via the EMCC-plume system discussed above; vertical fluxes across the seasonal pycnocline; and, winter convection, which supplies the standing stock of nutrients that fuels the spring phytoplankton bloom (Townsend, 1991). Additional vertical nutrient fluxes throughout much of the year in offshore waters are driven by processes associated with internal waves (Brickley, 2000).

Recent analyses have shown that nutrients that enter the eastern Gulf of Maine at the surface via Scotian Shelf Water may be as important to production as those that enter via the deep Slope Water that comes through the Northeast Channel (Townsend, 1997; 1998). This is because although the gross influx via Slope Water is much greater, only about 23% of that nutrient load reaches the surface layer (euphotic zone) where it becomes available to phytoplankton. Box model analyses have suggested that the influx into the Gulf of new nitrogen at depth through the Northeast Channel cannot sustain by itself all the observed primary production, because most of that nutrient load exits the Gulf in the intermediate water layer before being made available to the primary producers in surface waters. Nutrient fluxes from intermediate waters, where there is a primary nitrite maximum (Holligan et al., 1984), suggestive of localized nitrification, likely provide the surface water nutrients necessary to support the relatively high rates of primary production in the Gulf of Maine (Townsend, 1998).

Source water nutrient concentrations ultimately control the absolute amount of new primary production that can be supported in the Gulf of Maine. We discussed earlier how differences in nutrient loads throughout the region can be related with the type of bottom water; that is, nutrient concentrations are greatest for Warm Slope Water, which carry nitrate concentrations $>23 \mu\text{M}$ and least for cold Labrador Slope Water, which has on the order of $15\text{-}16 \mu\text{M}$ (e.g., see Fig. 5.10). Silicate concentrations in both water masses, on the other hand, are on

the order of 10-14 μM , which means that nitrate concentrations are greater than silicate. This excess of nitrate over silicate becomes important in determining the species composition of the phytoplankton. Because diatoms, which dominate the spring bloom period, take up nitrate and silicate in roughly equal proportions, silicate, not nitrate, would limit the spring bloom in the Gulf of Maine. On the other hand, nearer the coast and the influence of riverine sources of silicate, which can exceed 200 μM (Schoudel, 1996), the bloom would be limited first by nitrate. A survey conducted just following the Gulf of Maine spring bloom in 2000 showed that the surface waters in the offshore Gulf were depleted in silicate relative to nitrate, but that adjacent to the coast, silicate concentrations remained higher than nitrate (Fig. 5.14).

The spring phytoplankton bloom – in the Gulf of Maine as well as throughout the northwest Atlantic shelf region – is one of the most important biological oceanographic events of the year. The annual bloom in temperate and high latitude continental shelf waters is usually the time of highest concentrations of particulate organic matter (phytoplankton cells and aggregates) as well as dissolved organic material, and may represent more than half the total input of organic matter to deep water and the benthos (Parsons et al., 1984; Smetacek et al., 1978; Smetacek, 1980; Peinert et al., 1982). The spring bloom throughout the northwest Atlantic shelf region begins first in shallow inshore areas (Hitchcock and Smayda, 1977; Townsend, 1984; Townsend and Spinrad, 1986), when the critical depth (i.e., Sverdrup, 1953) exceeds the bottom depth, which can happen in winter. In deeper offshore waters the spring bloom is fueled primarily by nutrients mixed upward by winter convection. The degree of winter convective mixing likely varies among years (Brown and Beardsley, 1978), which could produce interannual variations in the concentrations of inorganic nutrients available in surface waters in spring, as could variations in the relative proportions of Labrador Slope Water and Warm Slope Water residing in the Gulf, as we discussed earlier. Factors that control the dynamics of the spring phytoplankton bloom have been recognized in a descriptive sense for decades (Gran and Braarud, 1935; Bigelow et al., 1940; Sverdrup, 1953), and in the Gulf of Maine, the possible causes of that variability have been discussed in greater detail and in the context of more recent field observations (Townsend and Spinrad, 1986; Townsend et al., 1992b, 1994b). Current thinking holds that oceanographic conditions for the initiation and evolution of the spring phytoplankton bloom in the deeper offshore waters of the Gulf will conform to one of three scenarios:

First, the bloom may be set up according to the classical Sverdrup (1953) model, whereby a thermocline develops in spring creating a shoaling upper mixed layer, which in conjunction with deepening light penetration in spring, reaches a critical light intensity in the upper layer and net planktonic production commences. Riley (1957; 1967) suggested that the value of the critical light intensity that triggers the bloom is reached when the depth-averaged solar irradiance within the mixed layer increases to ca. 20.9 W m^{-2} ; this has been corroborated by a number of subsequent reports from around the world (Gieskes and Kraay, 1975; Pingree et al., 1976; Hitchcock and Smayda, 1977) including studies in the Gulf of Maine (Townsend and Spinrad, 1986; Townsend et al., 1992b, 1994b).

Second, in a more novel scenario, it has been shown that the spring bloom in the Gulf of Maine may develop in the absence of any vertical water column stability at all (Townsend et al., 1992b, 1994b). That is, the spring bloom can begin following the winter period of convective mixing, and prior to the vernal development of vertical water column stability, provided that wind speeds are below a certain, predictable threshold, which in the Gulf of Maine, is about 20

kts (Townsend et al., 1994b). The crux of this argument is that although vertically well mixed (little if any density differences with depth), the water column may not necessarily be actively mixing. In such cases, if the growth rates of the phytoplankton exceed losses from sinking or grazing, bloom production may commence. In addition, the bloom may not exhaust the supply of nutrients prior to the development of the seasonal thermocline. Rather, there may be several spring bloom pulses, each interrupted by self-shading light limitation or vertical mixing events. Eventually the seasonal thermocline develops, isolating the surface layers from the deep water nutrient source, and nutrient exhaustion curtails bloom production. The possibility of a succession of episodic blooms means that the spring bloom period may be significantly more productive, result in more export production, and be more important to higher trophic level production than has been generally assumed.

Third, the spring bloom in the eastern Gulf of Maine and on the Nova Scotian Shelf may result from the presence of Scotian Shelf Water. As those cold (0-3°C), low salinity waters flow to the southwest, with the major portion flowing into the Gulf of Maine and some crossing the Northeast Channel to Georges Bank, they appear to bring with them sufficient buoyancy and inorganic nutrients such that, depending on solar irradiance levels (e.g., weather), initiation of a phytoplankton bloom can occur. The potential for highly efficient benthic-pelagic coupling of a phytoplankton bloom in such cold water is intriguing (e.g., Rivkin et al., 1996). For example, Pomeroy and Deibel (1986) showed that during early spring blooms in waters near 0°C off the coast of Newfoundland, the metabolism of heterotrophic activity is slowed greatly relative to the rate of photosynthesis by autotrophs, thus allowing fresh organic material to be delivered to the benthos relatively unrespired. Townsend and Cammen (1988) argued that such early blooms in cold waters could benefit recruitment of juvenile demersal fishes in the Gulf of Maine region by stimulating benthic productivity prior to the arrival of newly metamorphosed juveniles in late spring.

An example of an early phytoplankton bloom associated with Scotian Shelf Water was observed for a station sampled on 24 March 1997 over the Northeast Channel area of the Gulf where high phytoplankton chlorophyll levels were associated with cold, fresh Scotian Shelf Water that provided a stable water column (Townsend and Thomas, 2001; Fig. 5.15); also, a similar phytoplankton bloom is clearly visible in satellite ocean color images of Scotian Shelf Waters in March of both 1998 and 1999 (Fig. 5.11). Recent work in the Georges Bank-Gulf of Maine area has documented significant episodic fluxes of Scotian Shelf Water (SSW) off the Nova Scotian continental shelf edge to Georges Bank (Bisagni et al., 1996) and into the Gulf of Maine (Mountain and Manning, 1996). Bisagni et al. (1996) noted that the occurrence of SSW on Georges Bank was not related to upstream river discharge, and that the processes controlling its movements onto Georges Bank (and into the Gulf of Maine) remain unknown. Nevertheless, as the numbers of satellite and shipboard observations accumulate, we are seeing that SSW often is advected across isobaths to Georges Bank and spreads across the interior of the Gulf of Maine (Mountain and Manning, 1996).

The Bay of Fundy has not been as well studied over the years as the Gulf of Maine proper. It is an unusual marine system, in that it is the site of the world's highest tides, which can exceed 16 m (during spring tides) in the upper reaches of Minas Basin. Gordon and Baretta (1982) have described the system in detail. The upper Bay has extensive tidal flats and salt marshes, and its high benthic productivity supports large populations of sea birds. But, the swift

tidal currents in the inner Bay keep the waters vertically well mixed and highly turbid, creating a light-limited environment that impedes planktonic primary production (Hargrave et al., 1983). Phytoplankton production is roughly equal to that of the benthic macroalgae, which together are roughly equal to the primary production of detrital material in the extensive salt marshes in the upper Bay (Gordon et al., 1985; Keizer and Gordon, 1985). An exception is the waters over Grand Manan Basin, inside the mouth of the Bay (Fig. 5.1), which appear to be much more productive in terms of planktonic phytoplankton. The waters there are sufficiently deep (and hence surface waters are sufficiently removed from the influence of tidal mixing) that thermal stratification can develop in the warmer months; this offshore stratified water at the mouth of the Bay of Fundy can be seen as a region of warmer surface temperature in the satellite image in Figure 5.13. The elevated primary productivity of the deeper waters at the mouth of the Bay of Fundy most likely results from upwelling of nutrients on the Nova Scotian side of the Bay, as discussed earlier, and because tidal mixing of the deeper waters promotes the development of a shallow pycnocline and nutricline that are shallower than the critical depth (Townsend et al., 2001). This part of the Gulf of Maine is important to seasonal aggregations of migratory pelagic fishes, especially herring, and its prey, *Calanus finmarchicus*, and are an important feeding ground for several species of great whales (Mate et al., 1997; Kenney et al., 2001).

3.3. Georges Bank

Georges Bank, a shallow submarine bank sitting at the mouth of the Gulf of Maine (Fig. 5.1), is one of the most prominent features of the northwest Atlantic continental shelf region. It measures about 150 km by 200 km, rising steeply from greater than 300 m depth beyond the northern flank, whereas the southern half of the Bank has a gentler slope, dropping off more gradually to the shelf edge and slope. The Bank is dominated by tidal mixing currents throughout most of its area, being most pronounced in the central shallow region on the top of the Bank, inside the 60 m isobath, where tides keep the water column vertically well mixed throughout the year. The general pattern of surface currents in the Bank region has been studied since the time of Bigelow (1927) who first described the general clockwise (anticyclonic) flow around the Bank. Loder (1980), Butman et al. (1982), Butman and Beardsley (1987), Limeburner and Beardsley (1996) and others have provided much more detail, showing the tidally-rectified “jet” current along the steep northern flank of the Bank, the seasonal increase in residual current speeds around the remainder of the Bank, from a seasonal low of 4-6 cm s⁻¹ in February and March to 11-18 cm s⁻¹ in May and June, and the closed, but “leaky” recirculation of the flow at the southwestern edge. These studies formed the basis for the subsequent development of a number of numerical circulation models which have been used to simulate the main features of the physical oceanography of the Bank region and thus provide important additional insights (Lynch and Namie, 1993; Namie, 1996; Chen et al., 2001).

Our present understanding of the oceanography of Georges Bank waters has benefitted from the recent results of a concerted study of the Bank’s physical and biological oceanography (the GLOBEC program [Global Ecosystem Dynamics]), much of which is presented in Wiebe and Beardsley (1996) and Wiebe et al. (2001). But well before these modern studies the Bank had been known for its high rates of biological productivity and bountiful fisheries (reviewed in Backus, 1987). Primary productivity of Georges Bank is thought to be among the highest of any

continental shelf sea, with rates reported to exceed $400 \text{ gC m}^{-2} \text{ y}^{-1}$ in the central portion of the Bank (O'Reilly et al., 1987). The production cycle is highly seasonal, marked by a pronounced late winter-early spring phytoplankton blooms (Riley, 1941; Walsh et al., 1987; Cura, 1987; Townsend and Pettigrew, 1997; Townsend and Thomas, 2001; 2002).

The spring bloom on Georges Bank can begin as early as January (Townsend and Thomas, 2001; 2002) over the central shallow (<60 m) regions, when bathymetry determines the base of the upper mixed layer (Townsend et al., 1994b). The bloom is triggered once the critical depth exceeds the water depth, which can occur over the shallow portions of the Bank even during winter under conditions of reduced cloud cover (Riley, 1941). The winter-spring bloom on Georges Bank is dominated by diatoms (Cura, 1987; Kemper, 2000; Townsend and Thomas, 2001; 2002) and thus depleted silicate appears to limit bloom production as early as February when surface concentrations approach typical values reported in the literature for diatom half-saturation constants ($2\text{-}4 \mu\text{M Si(OH)}_4$; Paasch, 1973). This is illustrated in Figure 5.16, where comparisons of nitrate+nitrite and silicate in February show a greater depletion of silicate at most stations on the top of the Bank. It is only after silicate had already become depleted that dissolved inorganic nitrogen concentrations are reduced to levels that would limit phytoplankton production, which usually occurs in April (Townsend and Thomas, 2001; 2002). For the remainder of the year, primary production is thought to be fueled largely by recycled nitrogen (Horne et al., 1989).

Kemper (2000) and Townsend and Thomas (2002) showed that the springtime phytoplankton community composition on Georges Bank undergoes a species shift from nearly complete dominance of diatoms during the winter-spring bloom, to mixed diatoms and flagellates, and only very low densities of diatoms in May and June. Cura (1987) had noted the increasing importance of dinoflagellates in early summer. The phenomenon of a springtime decrease in diatom densities, reaching a minimum in May, is the apparent result of silicate depletion over the top of the Bank prior to depletion of nitrogen. As already discussed, the source waters available for mixing with waters on top of Georges Bank have approximately $4\text{-}6 \mu\text{M}$ greater concentrations of inorganic nitrogen than silicate (Townsend and Thomas, 2001). Even in January, Townsend and Thomas (2002) observed near zero silicate concentrations with corresponding dissolved inorganic nitrogen values that were about $5 \mu\text{M}$ greater. In May, and especially in June, they saw the silicate levels begin to increase to as high as $6 \mu\text{M}$ as a result of silicate regeneration (Townsend and Thomas, 2002).

During the six-month winter-spring period from January to June of 1999, Townsend and Thomas (2002) observed a steady increase in overall plankton biomass and an increase in planktonic food quality, as indicated by lower particulate C:N ratios. Much of this early summer plankton community would be dependent on recycled nutrients, especially nitrogen; in keeping with this scenario, Townsend and Thomas (2002) showed that the heterotrophic component of plankton increases in May and June, possibly facilitating the recycling of nitrogen, which drives the majority of planktonic primary production following the winter-spring bloom.

The foregoing discussion illustrates how source water nutrients are important to the biological oceanography of Georges Bank because of nutrient ratios (e.g., N : Si); but perhaps even more important is the nature of how those source water nutrients make their way onto the Bank and become mixed throughout its area. Cross-isobath, cross-frontal mixing and nutrient injections onto Georges Bank appear to be most important along the Northern Flank of the Bank

(Pastuszak et al., 1982; Townsend and Pettigrew, 1997; Houghton and Ho, 2001) where nutrient-rich slope water resides nearby, having entered Georges Basin via the Northeast Channel (Fig. 5.1). A schematic diagram that illustrates this conceptual framework of surface circulation, nutrient fluxes, new and recycled primary production, and secondary production of zooplankton might look like that presented in Figure 5.17. This figure illustrates how deep water nutrients enter Georges Basin via the Northeast Channel and become mixed up onto the Bank. The noticeably greater flux of nutrients to the northern edge of the Bank than elsewhere around its periphery (Pastuszak et al., 1982; Townsend and Pettigrew, 1997) often leads to greater phytoplankton biomass accumulations on the Northeast Peak (Cura, 1987), where the “jet” current on the northern flank spreads out (the widening light gray area in Fig. 5.17). Nutrient fluxes across the remainder of the Bank’s edges, especially on the broad, gently sloping southern flank, have much farther to go to reach the shallower waters on the top of the Bank and much of the nutrient flux is utilized in a subsurface chlorophyll maximum layer before ever reaching the central Bank (Townsend and Pettigrew, 1997). Likewise, much of the nutrient flux on the northern edge of the Bank is also utilized by phytoplankton before dispersing across the top of the Bank; but, unlike the southern half of the Bank, it is manifested in phytoplankton distributed vertically throughout the shallow water column in the vicinity of the tidal front there (Townsend and Thomas, 2001; 2002). More importantly, it spreads across the northeastern portion of the Bank and circulates to the Southern Flank, fueling higher trophic level production along the way. That is, the residual clockwise circulation around Georges Bank carries nutrients and developing phytoplankton populations with it, thus promoting the observed phytoplankton biomass maximum on the Northeast Peak (Cura, 1987), and perhaps fueling zooplankton production within the vicinity of the tidal mixing front and farther downstream on the southern flank. In fact, the highest zooplankton biomass is, on average, observed on the southern half of Georges Bank (Davis, 1984), which is downstream from the presumed highest levels of new phytoplankton production (Fig. 5.17).

Higher rates of new primary production along the northern edges of Georges Bank, in the vicinity of tidal mixing fronts and upwelling (Houghton and Ho, 2001), were measured by Loder et al. (1992). They used ^{15}N tracer techniques to measure f-ratios (the percentage of primary production fueled by new nitrogen (NO_3) fluxes to that fueled by NO_3 plus recycled NH_4 ; Dugdale and Goering, 1967; Eppley and Peterson, 1979) along a transect that ran from deep waters just north of Georges Bank to the tidally well-mixed waters on the Northeast Peak. Loder et al. (1992) found f-ratios as high as 0.7 in regions where nitrate is mixed upward and onto the Bank. Inside the frontal areas, and in the central regions on top of the Bank, the f-ratios were on the order of 0.1 to 0.2. Thus, nitrate fluxes appear to support the nitrogen requirements of about 70% of primary production along the Bank’s edges, while recycled ammonium supports 80-90% of primary production on the Bank itself. This means that despite the high measured rates of primary production throughout the majority of the Bank’s area (O’Reilly et al., 1987) the particulate nitrogen so formed is principally recycled primary production. Thus, fluxes of new nitrogen (principally nitrate) delivered to the Bank from deeper waters around its edges appear to be too low to support significant levels of new primary production (c.f., Dugdale and Goering, 1967; Eppley and Peterson, 1979) across the entire Bank (Loder and Platt, 1985; O’Reilly et al., 1987; Walsh et al., 1987), prompting Townsend and Pettigrew (1997) to argue that secondary production is likely to be nitrogen limited. The phenomenon of limited exchange between newly

upwelled waters along the Bank edges and waters on the shallow central portion of the Bank can be seen in the satellite image of sea surface temperature in Figure 5.13. Note that the edges of the Bank are cooler and represent newly upwelled and tidally-mixed waters, while the central portions of the Bank, on the other hand, constitute a shallow warm pool, reflecting a relatively prolonged residence time there and limited exchanges with colder upwelled waters from the edges of the Bank. Limited lateral mixing across the Bank would constrain the flux of new nitrogen throughout its large area. Consequently, the center of Georges Bank is an area of predominantly recycled primary production, as diagramed in Figure 5.17, and an area where secondary production would be nitrogen-limited (Townsend and Pettigrew, 1997).

The idea of nitrogen limitation of secondary production on Georges Bank is supported by observations of anomalously low zooplankton production on the Bank when compared with the high rates of planktonic primary production (Sherman et al., 1987). Total zooplankton production (sum of microzooplankton and macrozooplankton) on the Bank has been estimated at 18% of phytoplankton production, while in waters of the adjacent Gulf of Maine it is 26% (Cohen and Grosslein, 1987). Part of the reason for this difference – nitrogen limitation of secondary production – is revealed in the temporal and spatial patterns of annual zooplankton abundance in each area. The annual cycle of copepods on Georges Bank exhibits a mid-spring peak, which follows the spring phytoplankton bloom, and then it abruptly declines to low summertime levels; this is in sharp contrast to the Gulf of Maine, where the decline in summer is far more gradual (Sherman et al., 1987). That is, zooplankton production on Georges Bank is unusually low in summer, and may be limited by the low post-bloom nitrogen concentrations, and flux rates, over most of the Bank's area. The same principle is evident in the spatial distribution of zooplankton on the Bank as well. Zooplankton abundance after the spring bloom is generally greatest on the Southern Flank (Davis, 1984), which is in keeping of the hypothesis of nitrogen limitation on top of the Bank and greater nitrate fluxes on the Northern Flank. That is, the maximum zooplankton densities are normally found downstream of the area of maximal nitrate fluxes onto the Bank and downstream of an area of high "new" phytoplankton production on the Northern Flank and Northeast Peak (Cura, 1987). Zooplankton on the Southern Flank, then, are less likely to be nitrogen limited, and are most likely a delivered product of new production upstream. Such a conceptual framework as diagramed in Figure 5.17 also fits with our understanding of fish spawning strategies on Georges Bank. Generally speaking, cod and haddock spawn on the northern and northeastern parts of the Bank, and the residual circulation then carries the developing larvae to the Southern Flank (Mountain and Schlitz, 1987) where zooplankton abundances are greatest, and secondary production is less nitrogen limited.

Despite the potential for nitrogen limitation on Georges Bank, the fact remains that it is a productive fisheries ecosystem. We might suggest the intriguing possibility that offsetting the efficiency-lowering effect of nitrogen limitation on secondary production, there is significant enhancement of recycled production and nitrification (microbial oxidation of ammonium to nitrate). The same reason that nitrogen is limiting to new primary production and secondary production – the isolation of most of the Bank's area from deep water nutrient sources – could make recycling of nutrients more efficient. Isolation from deep water sinks, and vigorous tidal mixing over much of the top of the Bank, would enable more efficient recycling and utilization of nitrogen and other nutrients before being lost to deep water or burial in sediments. For similar reasons, the relative isolation of central Georges Bank results in the early recharge of nutrients in

the Fall, not during winter convection as is the case in the deeper waters throughout the region. That is, nutrients appear to be replenished in late fall as new nutrients mix across the flanks and onto to the top of the Bank at a time when light levels limit primary production. Pastuszak et al. (1982) have shown that nitrate concentrations on Georges Bank increase to 90% of the maximum winter concentration by December, from summer levels of less than 2 μM to $>6 \mu\text{M}$; but from December to February, which is when water temperatures are cooling fastest, nitrate increases less than 1 μM (Fig. 5.18). It is this relatively early nutrient recharge in late fall that perhaps sets the stage for the early winter-spring phytoplankton blooms discussed above, extending significantly the production season to near year-round, and is thus very important to the overall biological productivity of the Bank. This hypothesis of late-fall and winter phytoplankton production is consistent with observed satellite ocean color imagery (Thomas et al., 2003).

The commercial fisheries in the Gulf of Maine-Georges Bank region have a long history dating back more than 200 years, but starting in the middle of the last century, stocks were experiencing significant reductions (Cohen and Langton, 1992; Sinclair, 1996). Landings in this region have not only declined severely in recent decades, with concomitant species flip-flops (Cohen and Langton, 1992; Sherman et al., 1987) but, with few exception, many important stocks remain at low levels (O'Bannon, 2002).

3.4. Southern New England and Middle Atlantic Bight

The continental shelf region between Georges Bank and Cape Hatteras forms a comma-shaped system that tapers in width and curves a full 90° in orientation from east-west off southern New England to north-south at Cape Hatteras. Included here are a number of sub-regions of interest: Nantucket Shoals, Long Island Sound, the New York Bight, and the Delaware, Narragansett and Chesapeake Bays. Much of the general oceanography of the Middle Atlantic Bight is already discussed in earlier sections of this review.

Nantucket Shoals and Long Island Sound, on the southern New England shelf, both exhibit high biological productivity, but for different reasons. The relatively shallow Nantucket Shoals (<50 m; Fig. 5.1) is tidally well mixed, and apparently awash in new nutrients that upwell in the apex of the Great South Channel of the Gulf of Maine. Satellite images of sea surface temperature show that this region is consistently colder than surrounding waters (e.g., Fig. 5.13), which is strongly suggestive of upwelling of Gulf of Maine waters. The South Channel Ocean Productivity Experiment (e.g., see Kenney and Wishner, 1995) reported results from surveys that included a few stations along transects from the southwestern Gulf of Maine onto Nantucket Shoals. As part of that program, Durbin et al. (1995) showed that during June, deep Gulf of Maine waters were upwelling onto the Shoals, as evidenced by colder, saltier waters with relatively high concentrations of inorganic nutrients ($\text{NO}_3 > 5 \mu\text{M}$). Only a handful of stations extended to the Shoals, however, and we are unaware of any published accounts of more detailed nutrient measurements in this area. Nonetheless, SeaWiFS ocean color satellite imagery show consistently high phytoplankton chlorophyll on the Shoals throughout the warmer months (Thomas et al., 2003).

Long Island Sound, situated between southern New England, USA, and Long Island is a shallow (average depth 20 m) temperate estuary with salinity ranging between 23 and 31

(Capriulo et al., 2002). The estuarine-like circulation has been reviewed by Bogden and O'Donnell (1998). Capriulo et al. (2002) have provided a general overview of the plankton of the Sound, which, they suggest should be considered an update of the earlier pioneering work of Gordon Riley and co-workers (e.g., Riley et al., 1956) who presented the first comprehensive overview study of the Sound's waters and benthos. The primary productivity is very high, approximately $470 \text{ gC m}^{-2} \text{ yr}^{-1}$ (Riley et al., 1956), much of which is driven by anthropogenic nutrient additions from the highly populated surrounding area.

On the other side of Long Island, the shelf waters of the New York Bight, in the northern portion of the Middle Atlantic Bight, are more biologically productive than the Gulf of Maine, but less productive than Georges Bank; that production derives from a combination of estuarine-like physical processes (from the Hudson River discharge), as well as cross-shelf interactions that promote nutrient fluxes and high rates of biological production (Falkowski et al., 1980). Much attention has been given to hypoxic conditions in past decades, which have been attributed to a combination of natural and anthropogenic causes (Pearce, 2000).

Continuing to the south, the continental shelf from Georges Bank to Cape Hatteras has a gentler slope running from the coast to the shelf edge. It is broadest in the north, becoming progressively narrower toward the south, and is shallower throughout its length than either the Nova Scotian Shelf or the Gulf of Maine. The residual circulation is dominated by an extension of the general equator-ward flow of shelf waters discussed earlier in this review, and is complicated by interactions with the shelf slope front as the shelf width constricts the flow. Two large estuarine systems dominate the nearshore coastal environment of the Middle Atlantic Bight, the Delaware and Chesapeake Bays. Biological productivity in each is high, and each has experienced new and emerging problems related to anthropogenic nutrient enrichment (see reviews in Sharp et al. (1982) and Magnien et al. (1992)).

Cross frontal mixing events between slope and shelf waters are important in nutrient fluxes in the Middle Atlantic Bight (Walsh et al., 1978; Marra et al., 1990). Walsh et al. (1978) showed that significant wind-driven upwelling from storm events can provide one third of the nitrate necessary to meet the primary production demand. They also reported that rates of primary production were an order of magnitude greater at a point 60-70 km offshore, in the vicinity of the shelf slope front, than on either side. This result is similar to that reported on the Scotian Shelf, as discussed above, and is consistent with the now-classic map of primary productivity produced by J. O'Reilly and his co-workers (Fig. 5.19; O'Reilly and Busch, 1984; O'Reilly et al., 1987; O'Reilly and Zetlin, 1998). The species composition and abundance of the phytoplankton community has been described for the northwest Atlantic shelf region in a series of papers by Marshall and his coworkers (Marshall and Cohn, 1983; Marshall, 1982; 1984; Marshall and Alden, 1993); they noted in general that highest cell densities were distributed in inshore areas, along the shelf break and on Georges Bank.

Yoder et al. (2001) analyzed coastal zone color scanner data available from 1987 to 1989 in an effort to characterize the temporal and spatial variability in near surface chlorophyll concentrations throughout the region. The Middle Atlantic Bight was divided into ten primary zones and the region was examined for coherency and signals across time and space. They found that the area is dominated by a clear, highly coherent seasonal signal with relatively low along-shelf, and high cross-shelf, variability. The region around Nantucket Shoals and Georges Bank are most variable and there is a trend of higher chlorophyll concentrations inshore than

offshore. The timing of the winter maximum and summer minimum concentrations varied slightly and there were secondary spring peaks on the outer shelf and slope of this region. They also found evidence of a high degree of short term variability superimposed on the general season signal, pointing to the importance of small scale events such as upwelling, storms, and fronts.

The general oceanography of the Middle Atlantic Bight shelf waters was reviewed extensively in Gross (1976). As we discussed earlier, this region was the focus of major research initiatives in the past, SEEP I and SEEP II (Shelf Edge Exchange Programs), which were concerned with the exchanges of carbon and other biogenic materials between the continental shelves and the deep ocean (Walsh et al., 1988; Biscaye et al., 1994). Much more continues to be learned about the interdisciplinary oceanography of the shelf waters of the Middle Atlantic Bight as a result of recent research initiatives, especially the Ocean Margins Program (e.g., see Bauer et al., 2002).

4. Summary

The northwest Atlantic continental shelf waters comprise a large geographical region within which we can identify several regional subsystems, all interconnected to some extent by an equatorward-flowing coastal current that has its origins as far away as the Labrador Sea. A wealth of scientific information on the oceanography of these waters has accumulated over the past century, making this region of the world ocean one of the best studied. But, at the same time, such a rich body of literature makes a brief review such as this a cursory one at best; for this, we apologize – indeed we could have, and would like to have, taken our discussions of the subregions into much greater detail. We can define the overall area as encompassed by a southern edge marked by a relatively abrupt transition off Cape Hatteras, where the Gulf Stream comes to within 30 km of shore. Shelf and slope waters are thus geographically constrained by the position of the Gulf Stream to the southeast, and, of course, the coast to the northwest. Within this set of boundaries, shelf and slope waters mix in complex ways both at the surface and at depth, and, because the main source of nutrients is offshore slope waters, which themselves can differ markedly in their nutrient loads, such mixing processes can be very important in setting levels of primary production. The nearshore, coastal environments throughout the region range from the deep, tidally energetic Bay of Fundy, to the shallow estuarine waters of Long Island Sound and the Chesapeake and Delaware Bays, which encompass ecosystem structures very different from one another. Each of these relatively large subunits is a shallow estuarine system that today is significantly affected by anthropogenic influences. Throughout the region, anthropogenic nutrient additions show no clear signs of abatement. On the other hand, the northwest Atlantic continental shelf, throughout its length, is apparently under the influence of a large scale atmospheric phenomenon – the North Atlantic Oscillation – which might be driven by climate change. The recent literature is beginning to report NAO effects that range from altered bottom water temperatures and altered rates of planktonic primary production to fluctuations in population sizes of expatriate copepods and, perhaps, commercial fish stocks. Whether driven by human actions or natural climatic forces, we know that commercial fisheries, especially on Georges Bank, are at historically low levels and have different species composition than was the case several decades ago. The thread that

ties all of these observations together is a wide open future filled with scientific research opportunities, from the most applied aspects of fisheries science and coastal pollution studies, to the most basic aspects of physical-biological coupling. Even though so much is known about this region, there remains much more to be discovered.

Acknowledgments

This review has drawn heavily on the many discussions we have had with the various researchers who have contributed so much to our understanding of the general oceanography of the waters of the northwest Atlantic continental shelf, and we apologize if we have in any way misrepresented their work or insights they have shared with us. In addition, we point out that a review such as this could easily be expanded fill a volume of its own, given the large body of published work that exists, and no doubt we have left out discussions of many significant studies; we trust our colleagues will understand this omission and forgive us. We thank J. Eberwine and K. Drinkwater for providing data presented in Figure 5.5. We gratefully acknowledge the assistance of R. Weatherbee for processing our satellite images and the constructive comments of R. Houghton and G. Perillo on an earlier draft of this paper.

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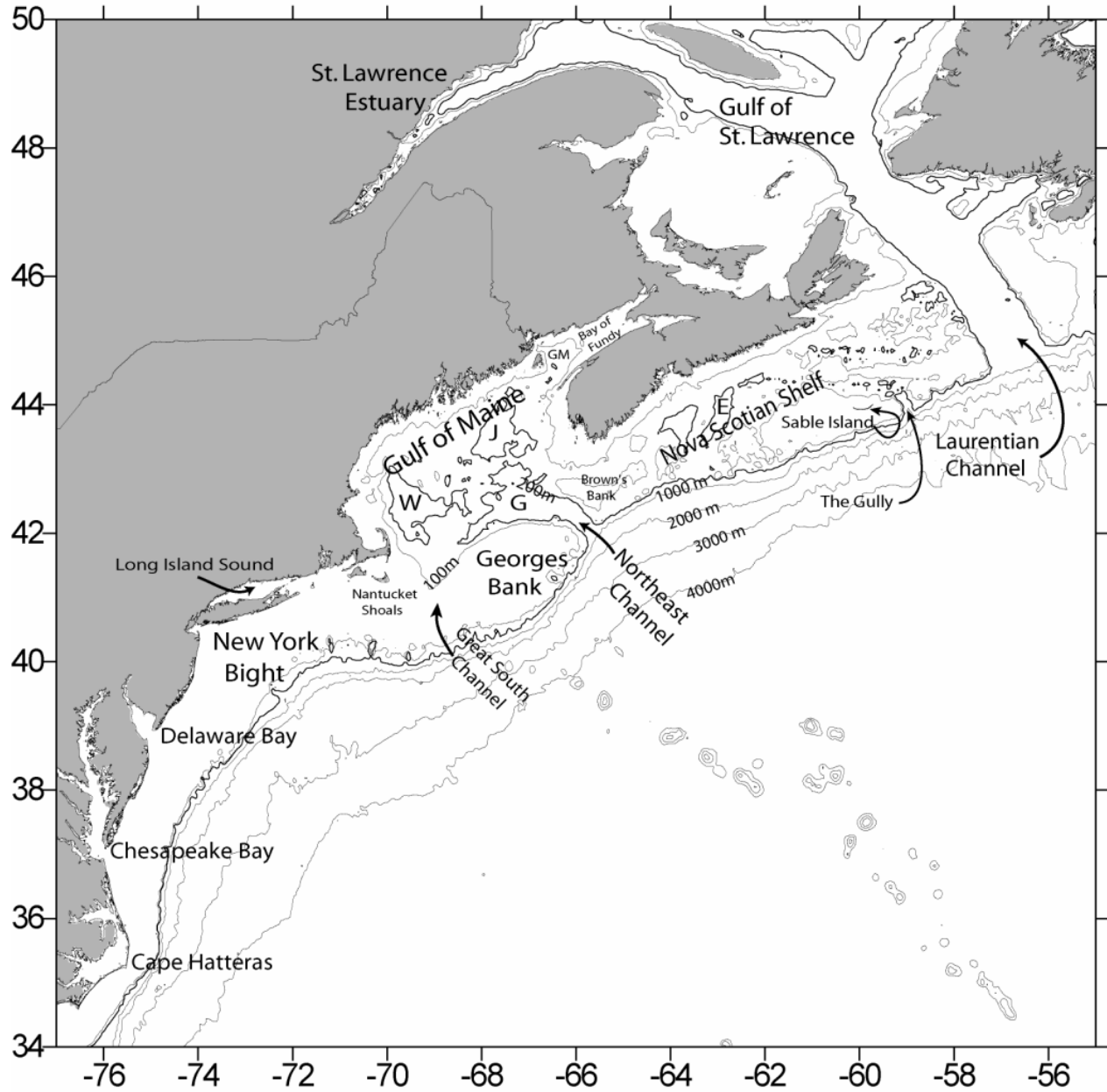


Figure 5.1. Map of the region with bathymetric contours at 100m, 200m (heavy line), 1000m, 2000m, 3000m and 4000m. Features referred to in the text are indicated. Basins in the Gulf of Maine and on the Nova Scotian Shelf are indicated as: W (Wilkinson Basin), J (Jordan Basin), G (Georges Basin) E (Emerald Basin) and GM (Grand Manan Basin).

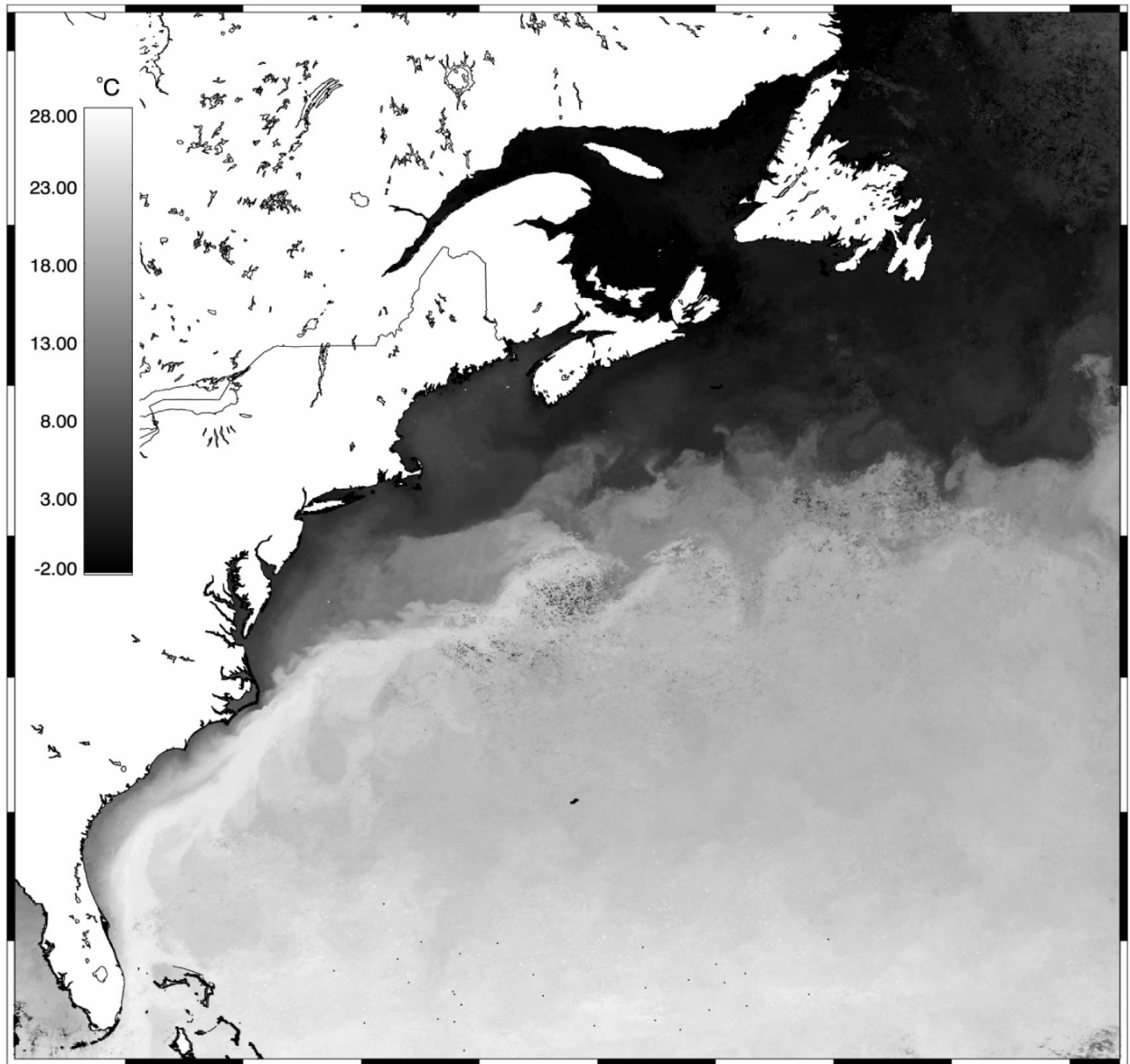


Figure 5.2. Composite satellite image of sea surface temperature for the period 10-28 February 2004, illustrating sharp temperature gradients separating the Gulf Stream waters in the south from colder waters of the study region. The image includes all cloud-free pixels from up to seven images per day measured by the Advanced Very High Resolution Radiometer (AVHRR) aboard a constellation of NOAA satellites.

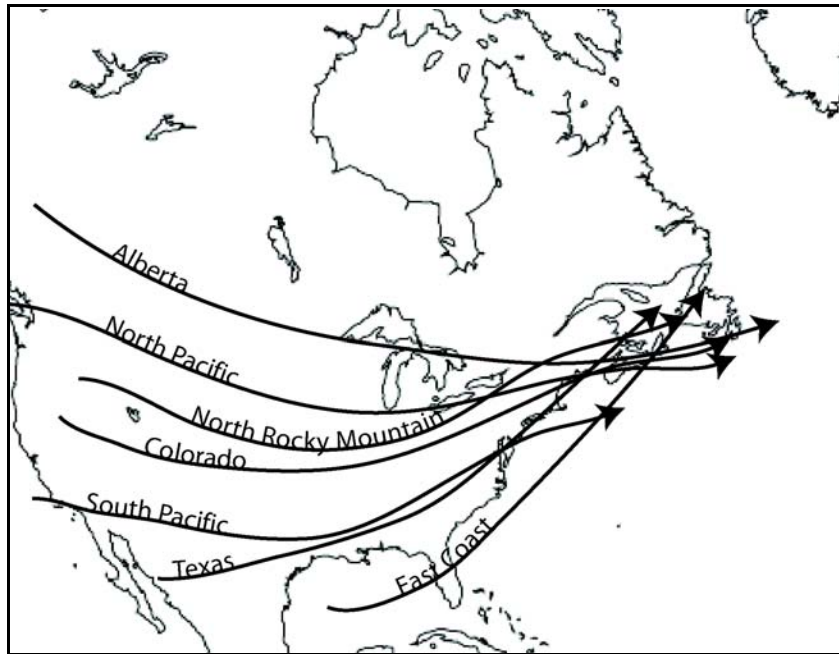


Figure 5.3. Major storm tracks across the North American continent, and their convergence in the study region.

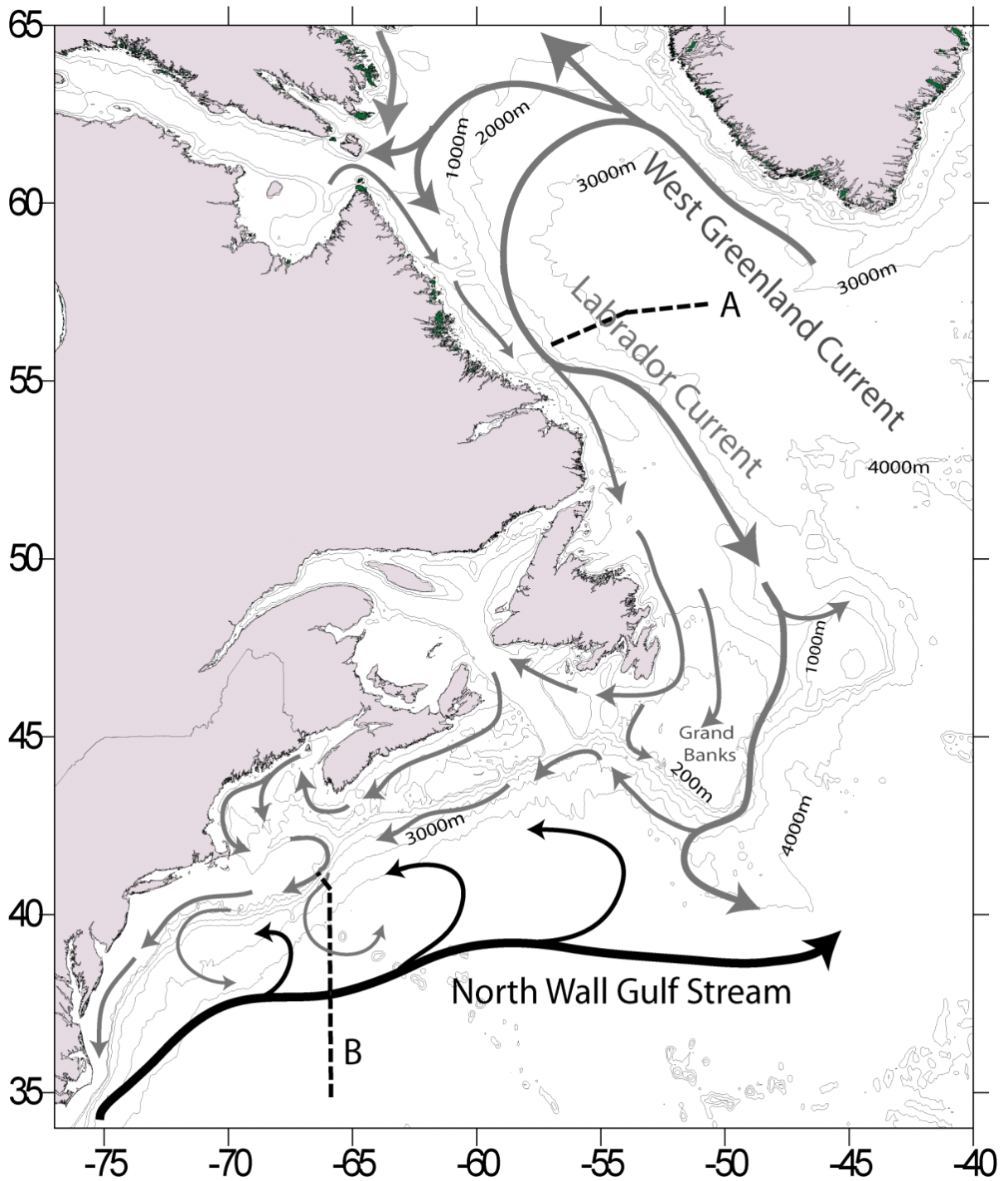


Figure 5.4. Major current systems in the region. The 4000m, 3000m, 2000m, 200m and 100m isobaths are indicated. Transects A and B are discussed in the text.

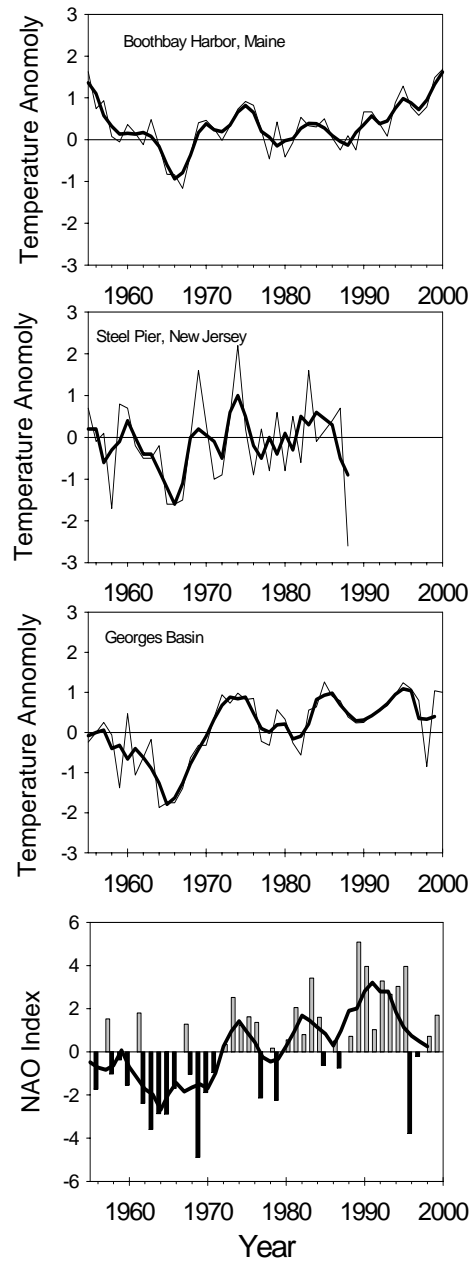


Figure 5.5.

Top 3

Panels: Time series of summertime surface temperature anomalies in the Gulf of Maine (Boothbay Harbor, Maine, top panel; data provided by the Maine Department of Marine Resources, West Boothbay Harbor, Maine); the New York Bight (Steel Pier, New Jersey, second panel; data provided by J. A. Eberwine, National Weather Service Forecast Office, Mt. Holly, New Jersey); and, bottom water temperature anomalies in Georges Basin (third panel; data provided by K. Drinkwater, Dept. Fisheries and Oceans, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada); annual data (thin lines) and 3-year running averages (thick lines) are given. Bottom panel: wintertime (Dec-Feb) North Atlantic Oscillation Index, as annual data (bars) and 5-year running average (line).

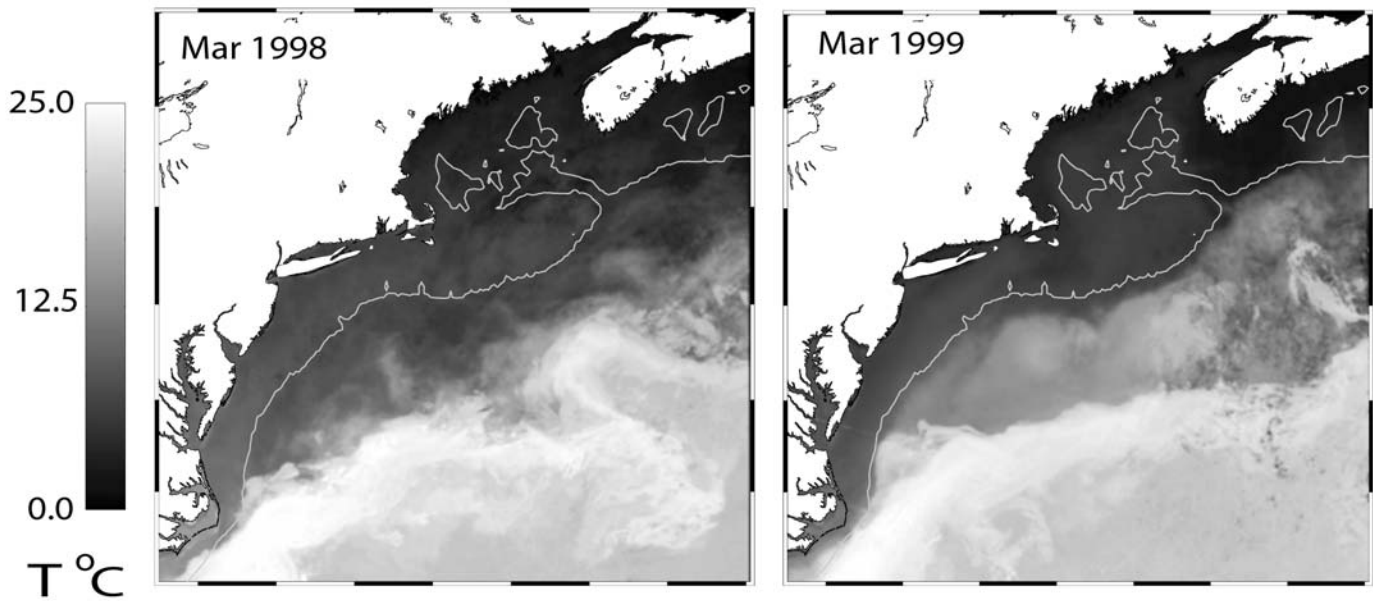


Figure 5.6. Satellite SST images from March 1998 (left panel) and March 1999 (right panel). The images show variable penetration of Labrador Slope Water between the shelf edge (200 m isobath is indicated) and Warm Slope Water associated with the north wall of the Gulf Stream. In 1998 (left), during an NAO low phase, cold Labrador Slope Water is adjacent to the shelf break, while in 1999 (right) Warm Slope Water is adjacent to the shelf break. Images are monthly composites of sea-surface temperature measured by the Advanced Very High Resolution Radiometer (AVHRR) aboard a constellation of NOAA satellites. All cloud-free pixels from up to seven images per day contribute to each composite.

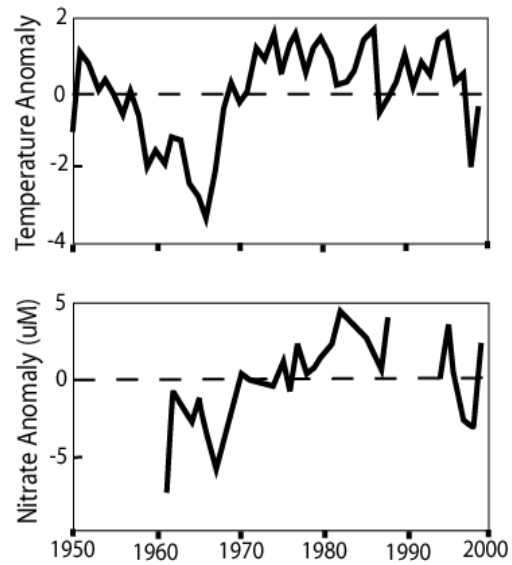


Figure 5.7. Time series of temperature and nitrate anomalies (μM) between depths of 100 to 200m on the central Nova Scotian Shelf (from Drinkwater et al., 2002).

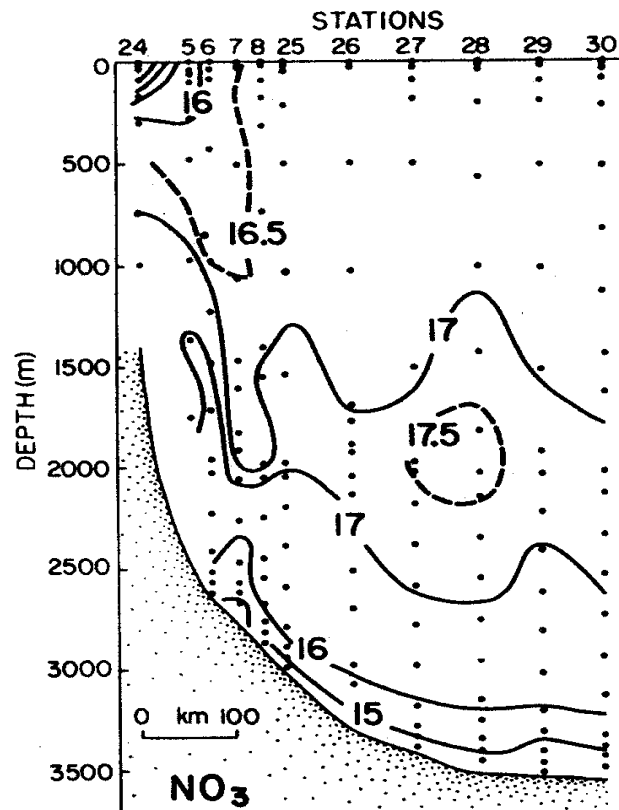


Figure 5.8. Concentrations of nitrate (μM) along a transect at approximately 55°N from the coast of Labrador (from Clarke and Coote, 1988). Corresponds to transect A in Figure 4.

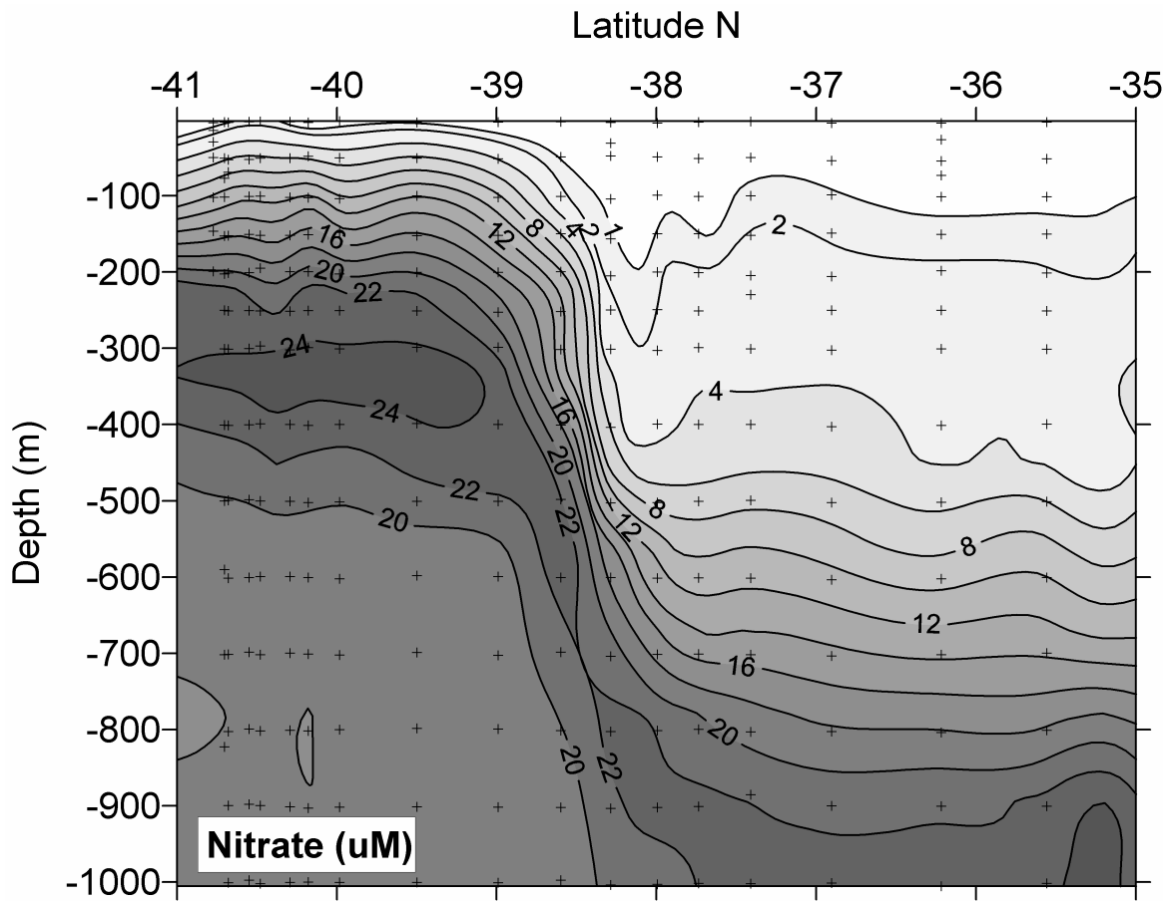


Figure 5.9. Concentrations of nitrate (NO_3 μM) along a transect from 41°N to 35°N along 66°W Longitude off the southeast edge of Georges Bank and into North Atlantic Central Water (transect B in Fig. 4). Figure is plotted using data from the World Ocean Circulation Experiment (WOCE) Hydrographic Program (WHP) at: <http://whpo.ucsd.edu/> for cruise A22 (15 Aug - 3 Sep 1997, R/V Knorr) described in Joyce et al., 2001. Cross hatches indicate sample depths.

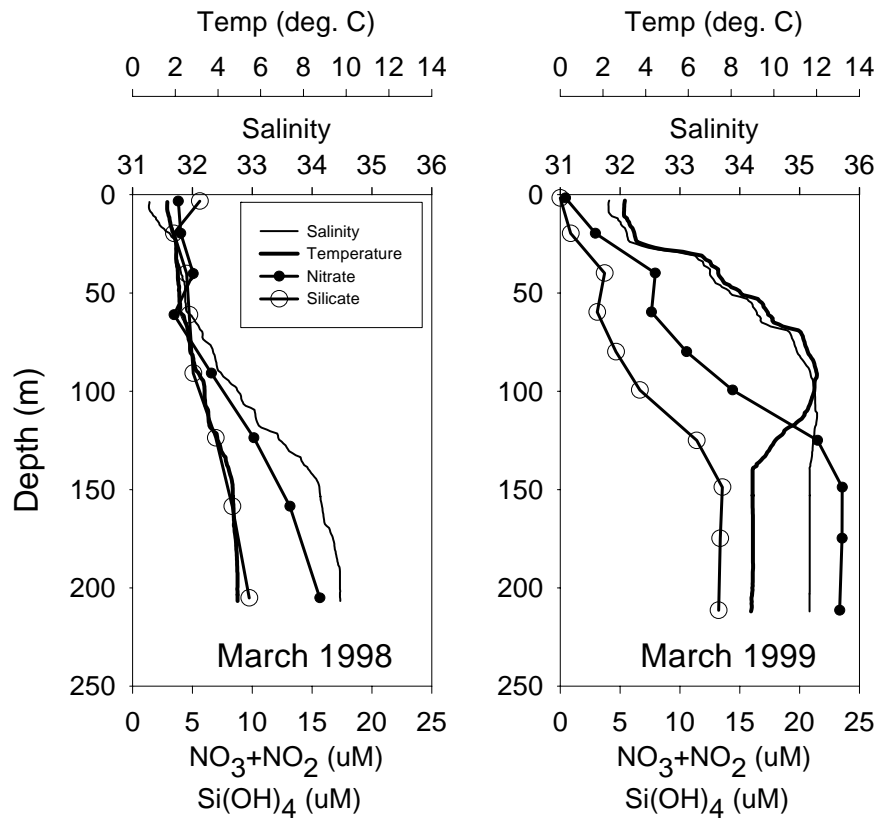


Figure 5.10. Vertical profiles of temperature, salinity, nitrate+nitrite and silicate in the Northeast Channel in March of 1998 and 1999 (Data from Townsend, unpublished).

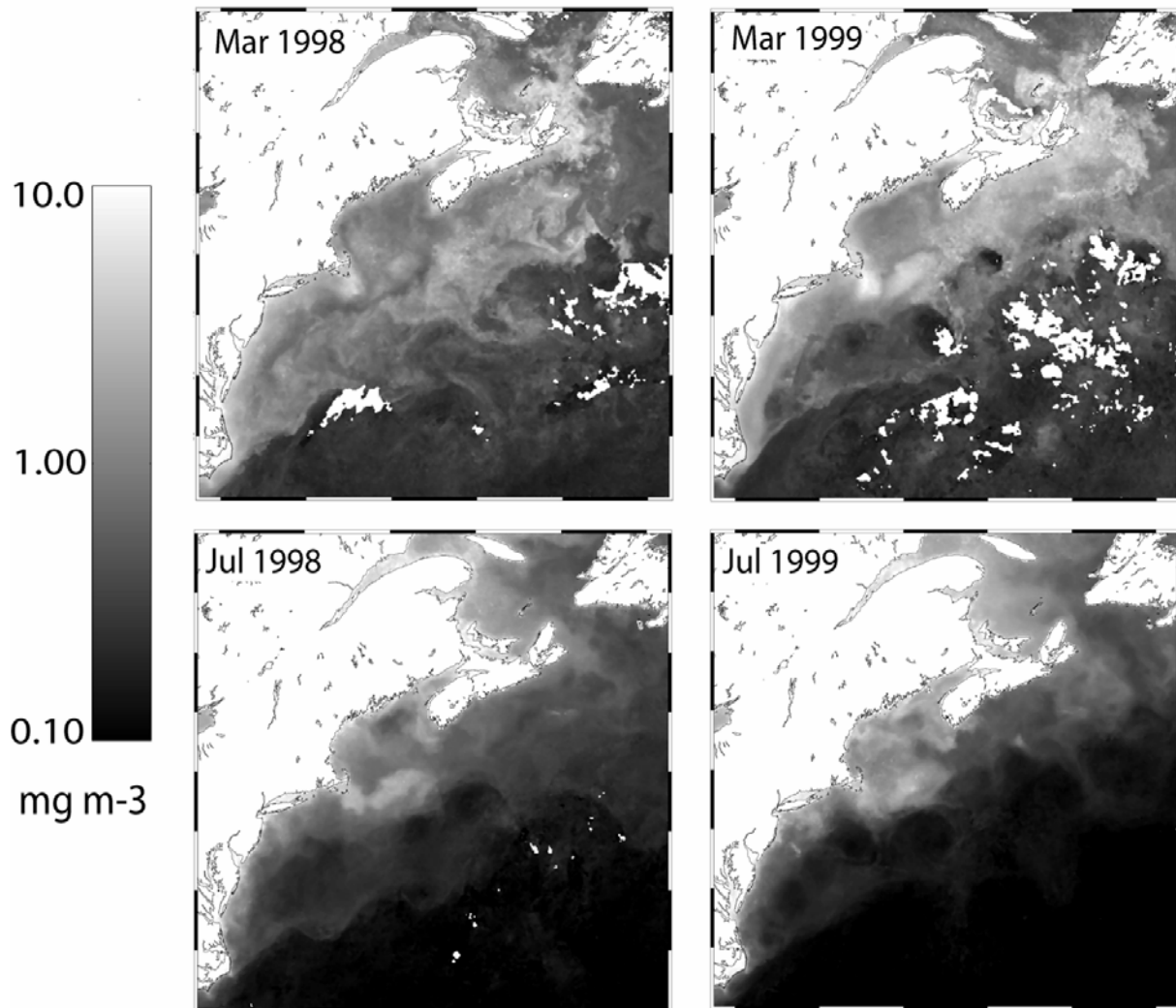


Figure 5.11. Monthly composites of surface chlorophyll concentrations measured by the Sea-viewing Wide Field-of-view Sensor (SeaWiFS satellite). All cloud-free pixels from 30 or 31 daily images during March and July of 1998 and 1999 contributed to each composite.

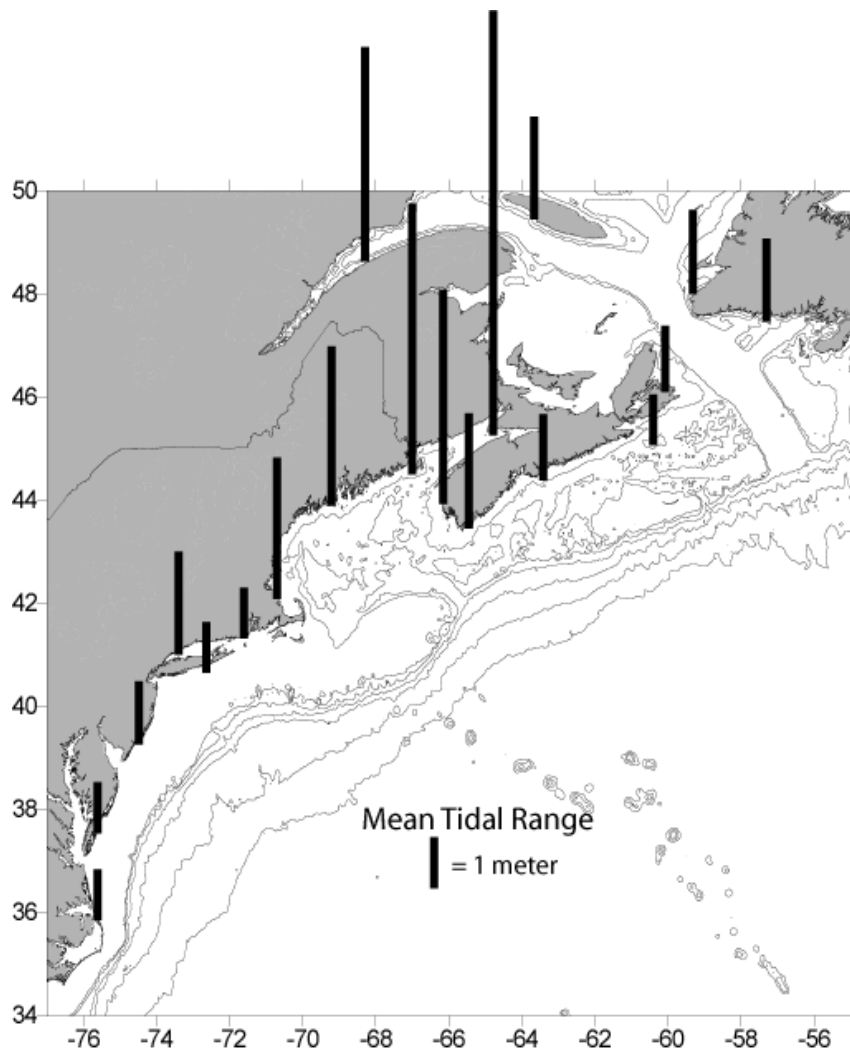


Figure 5.12. Schematic representation of tidal ranges throughout the region.

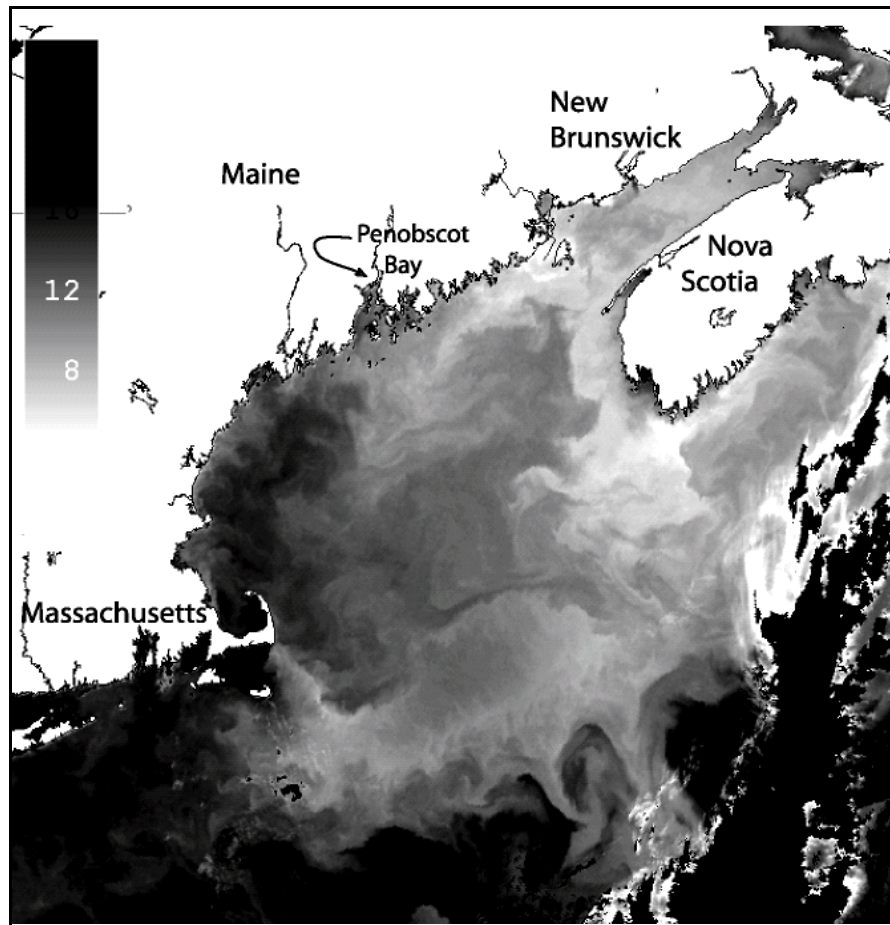


Figure 5.13. AVHRR satellite image of sea surface temperature in the Gulf of Maine region in June 1998 (year day 153). Locations referred to in the text are indicated.

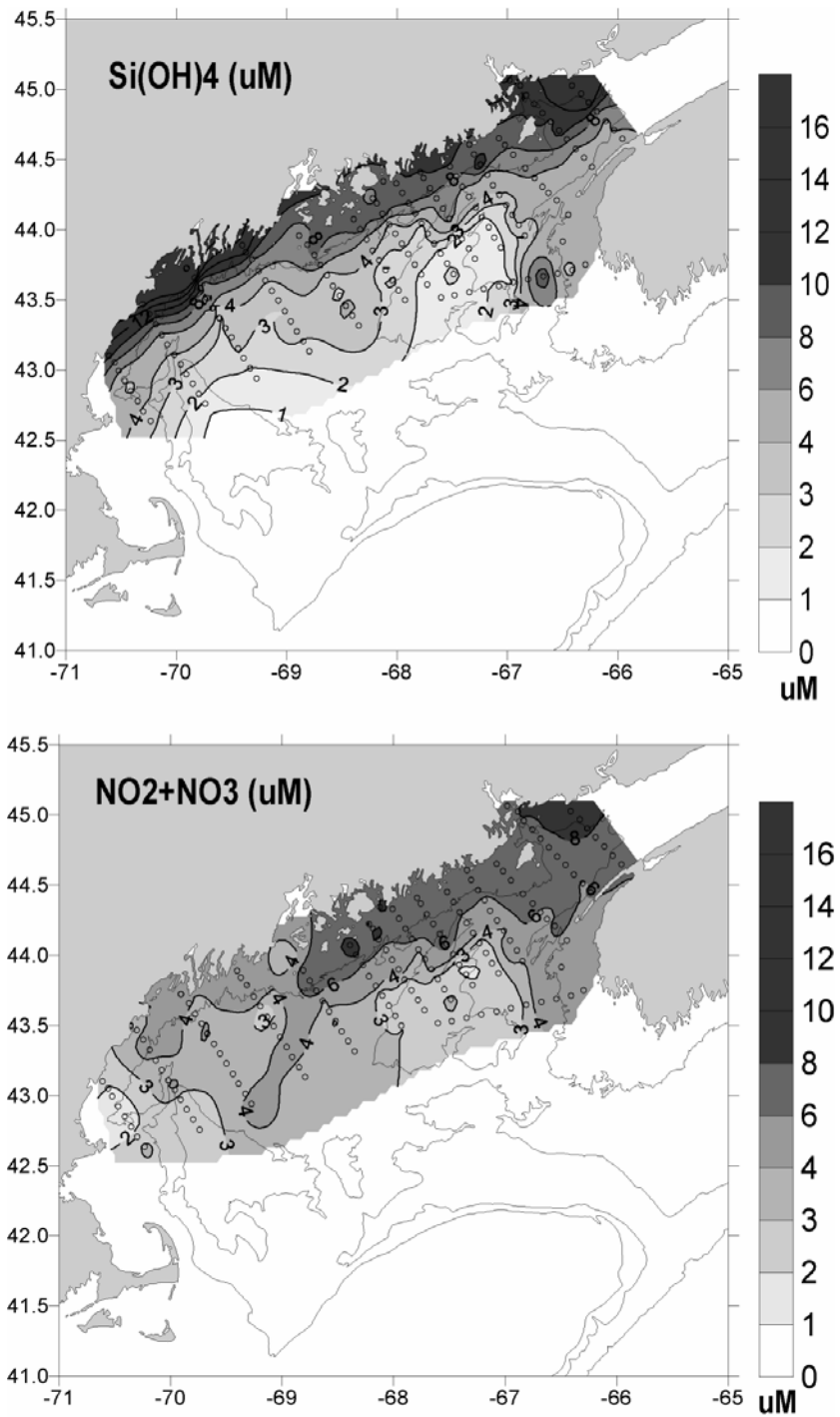


Figure 5.14. Surface water (2m) concentrations of Si(OH)₄ and NO₂+NO₃ in the Gulf of Maine in April-May of 2000 (Townsend, unpublished).

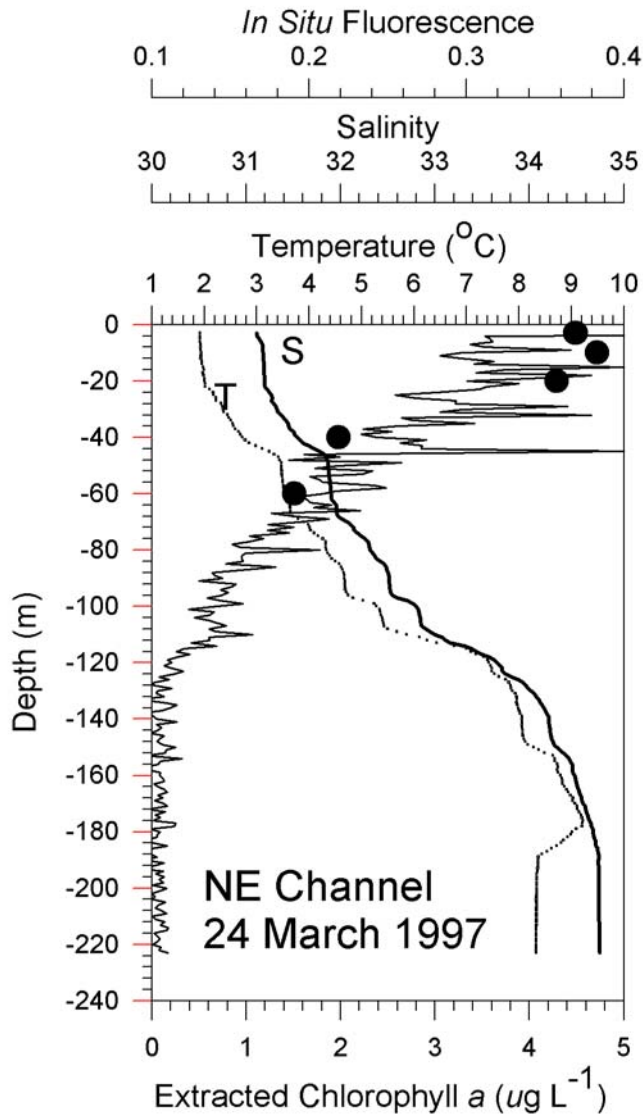


Figure 5.15. In situ phytoplankton chlorophyll fluorescence, extracted chlorophyll (•), and temperature and salinity at a station samples in the Northeast Channel of the Gulf of Maine on 24 March 1997 (from Townsend and Thomas, 2001). Note the bloom is in cold, fresh surface waters of Nova Scotian Shelf water origin.

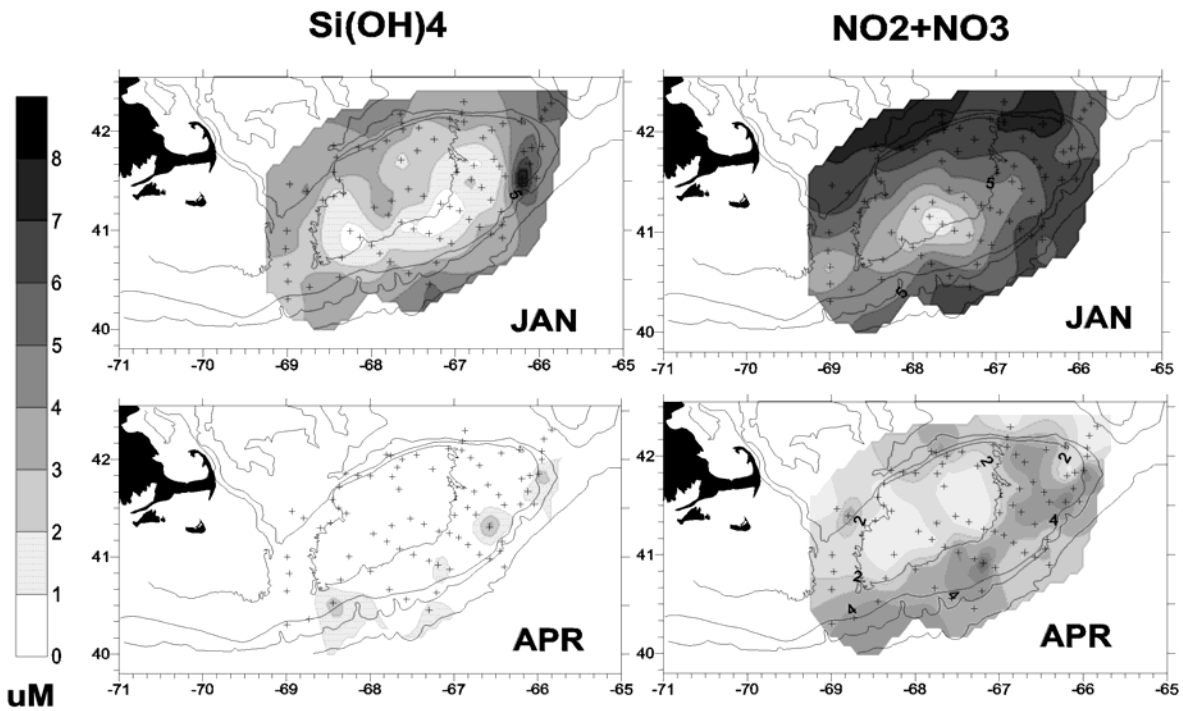


Figure 5.16. Surface water (2m) concentrations of Si(OH)_4 and $\text{NO}_2 + \text{NO}_3$ on Georges Bank in February and April of 1999 (from Townsend and Thomas, 2002).

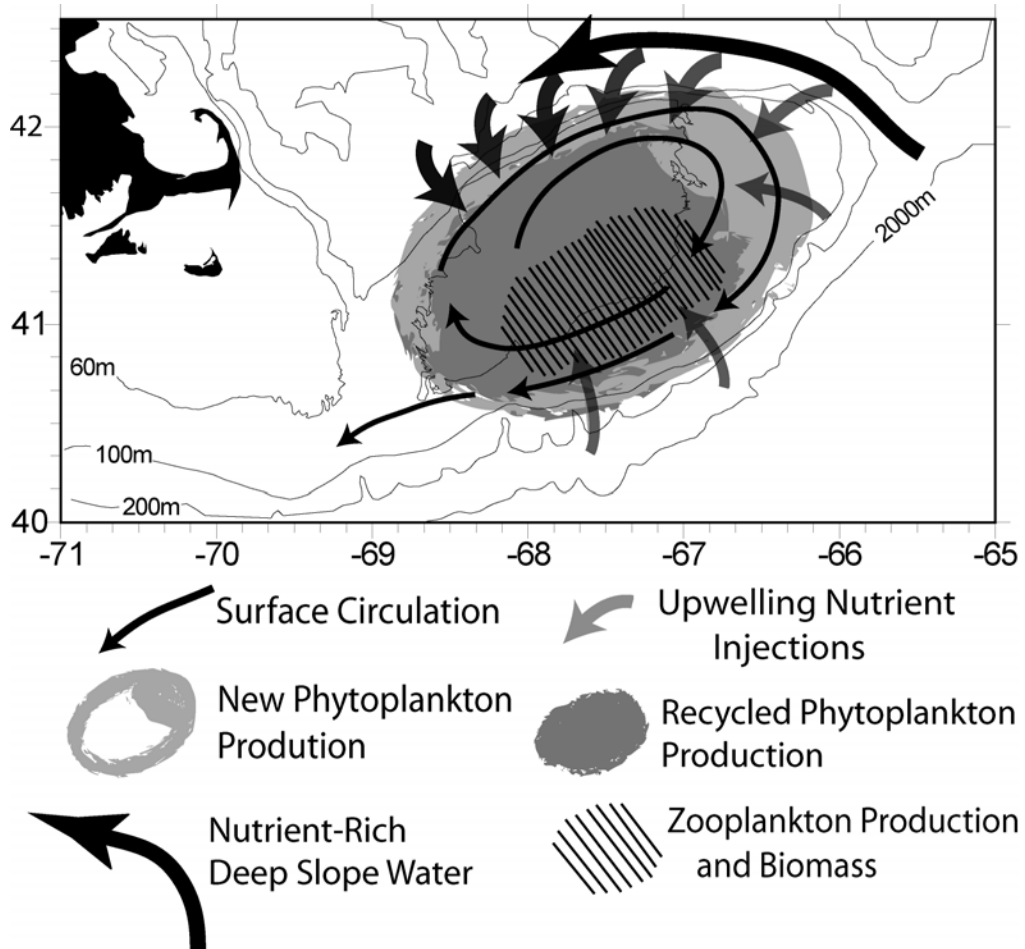


Figure 5.17. Schematic of surface circulation, flux of deep waters and nutrients, and resulting patterns of biological productivity on Georges Bank.

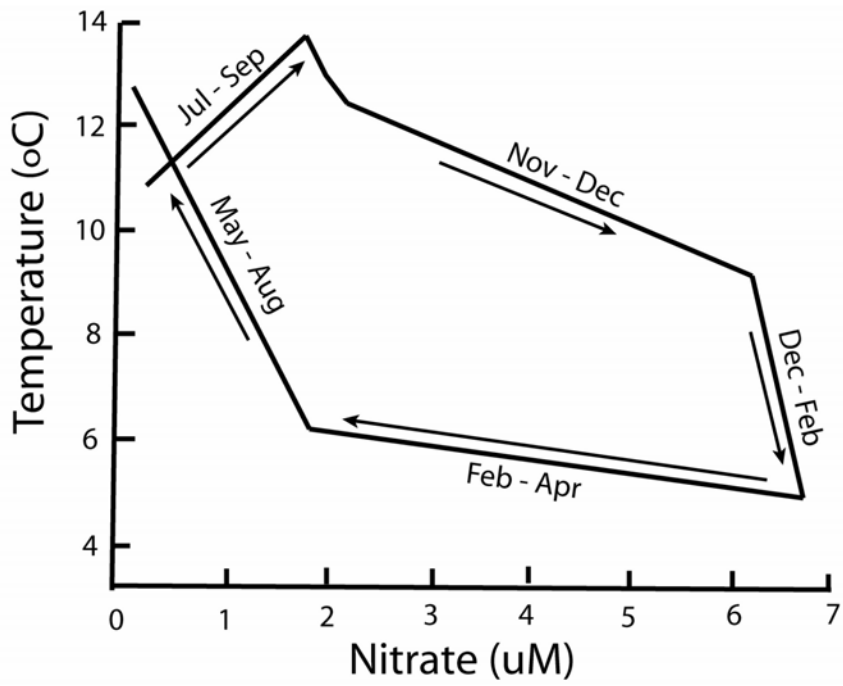


Figure 5.18. Annual cycle of temperature and nitrate on the central crest of Georges Bank, 1975-1976 (after Pastuszak et al., 1982).

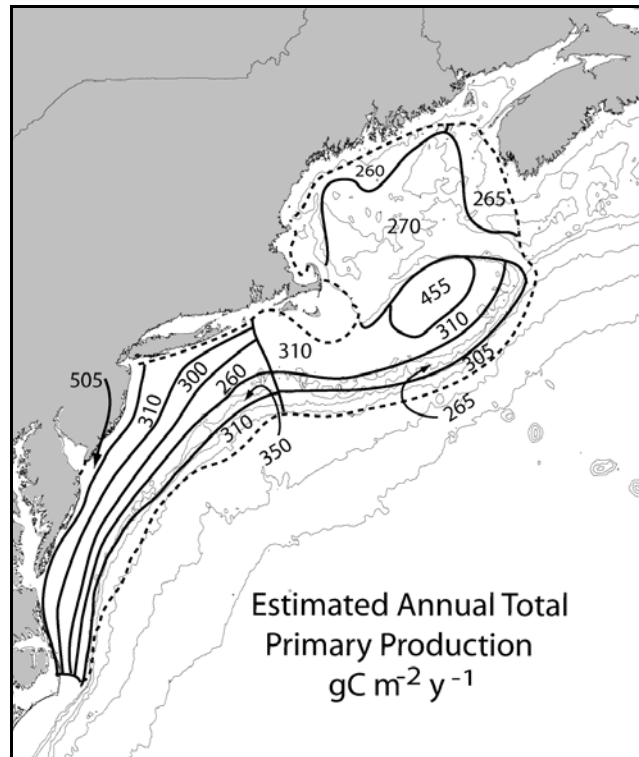


Figure 5.19. Annual primary production in the region (after O'Reilly et al., 1987).