Overrunning of shelf water in the southern Mid-Atlantic Bight

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Abstract

Analyses of two years (1992 and 1993) of high-resolution sea surface temperature satellite images of the southern Mid Atlantic Bight (MAB), showed that unusually extensive overhang of shelf water occurs episodically, and coherently over along shelf distances of several 100 km. These episodes are dubbed overrunning of the Slope Sea by shelf water. The overrunning volume has a “face” and a “back” (southern and northern limit). It transports substantial quantities of shelf water southward, and does not retreat onto the shelf, but eventually joins the western edge of the Gulf Stream in the vicinity of Chesapeake Bay. The combined analyses of satellite imagery and various in situ data further demonstrated that the shelf waters overrunning the Slope Sea were not mere surface features but reached depths between 40 and 60 m. Results confirm previous concepts on shelf circulation, shelf–slope exchange and fate of shelf water. They also shed new light on shelf water budget: overrunning of the Slope Sea and southwest transport by upper slope current constitutes an important conduit for shelf water transport. Winter storms move the shelf–slope front, and with it shelf water, offshore to distances ~10–40 km. The offshore displacement of shelf water can be related to the onshore veering of the Gulf Stream near Cape Hatteras, producing a blocking effect on the shelf circulation. Such a blocking effect of the southwestward flow of shelf water in the MAB appeared to be the reason for the overrunning of shelf water off New Jersey. In addition, the excess fresh water discharge from the St. Lawrence was also observed to be related to the overflow of shelf water off New Jersey.

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1. Introduction

The Mid-Atlantic Bight (MAB) extends from the southwest corner of Georges Bank to Cape Hatteras, North Carolina (Fig. 1) and has been extensively studied from the time of Bigelow (1933). Reviews of its physical oceanography can be found in Allen et al. (1983), Brink (1987) and in Epifanio and Garwine (2002). The MAB shelf topography is simple and smooth over its 800 km length, indented with canyons including the Hudson Canyon. The width of the shelf narrows from New York Bight (150 km) to Cape Hatteras (50 km). Wedged in between the shelf water and the Gulf Stream is the distinctive slope water mass named

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by Iselin (1936). The strip of ocean containing slope water is about 200 km wide and 1600 km long, stretching from Cape Hatteras to the Grand Banks, between the Gulf Stream and the continental shelf. It is called the “Slope Sea” by Csanady and Hamilton (1988).

The relatively cool and fresh shelf water is separated from the warmer and saltier continental slope water by the baroclinic shelf–slope front. The sharp shelf–slope frontal zone can be identified by its $T/S$ characteristics; usually an area enclosed between 34–35 psu and 6–13 °C. This thermohaline feature is discernible from around Georges Bank to Cape Hatteras and slopes downward and shoreward from the inner slope to where it intersects the bottom at about the 90 m isobath. The front is subject to a host of higher frequency perturbations apart from the seasonal changes which give it the highly convoluted surface appearance (Halliwell and Mooers, 1979).

The shelf water volume in the MAB is circumscribed by the coast and the shelf–slope front. Wright and Parker (1976) estimated that about half of the volume of shelf water from Cape Cod to Cape Hatteras lies in the surface wedge outside of the 100 m isobath and the shelf–slope front. Mountain (1991) in his analysis of the volume of shelf water in the MAB and Georges Bank determined the mean position of the shelf–slope front from satellite imagery. He suggested that changes in the location of the shelf–slope front at the surface, could account for approximately 50% of the variance in the shelf water volume. In addition, he observed that the offshore position of the front was highly variable and could account for 50–80% of the interannual changes.
in the volume of shelf water in the MAB. Also large interannual changes are observed in the volume of shelf water in the MAB ($2 \times 10^3 \text{ km}^3$). These have been related to the fresh water discharge from the St. Lawrence that advects into the Gulf of Maine and that later constitutes the major part of the inflow into the MAB. A number of processes contribute to the transfer of shelf water into the slope region in the MAB and constitute important mechanisms to transfer heat, salt and nutrients between the shelf and slope waters. This exchange and the mixing of shelf and slope waters is important for at least two reasons (Churchill et al., 1993). The first is the concern for the transport of contaminants to and from the continental shelf and the other is the transport of plankton to different hostile and hospitable environments. Knowledge of this exchange process is incomplete (Garfield and Evans, 1987). The movement of the shelf–slope front in the MAB plays an important part in the exchange of shelf water (Garvine et al., 1988). In simplified models, the front usually intersects the surface within 50 km offshore of the shelf break (200 m isobath in the MAB) and just inshore of the shelf edge at the bottom. Large surface displacements of 100–300 km are observed to take place (Halliwell and Mooers, 1979), but the bottom intersection of the front is found to vary only within 16 km of the shelf break (Wright, 1976).

Shelf Water is entrained from the western end of the Slope Sea eastward, along the northern boundary of the Gulf Stream (Ford et al., 1952; Kupferman and Garfield, 1977; Fisher, 1972; Gawarkiewicz et al., 1996; Churchill and Berger, 1998), reaching depths of about 100 m on the Gulf Stream side (Churchill et al., 1989). The entrainment into the warmer saline Gulf Stream can cause salt finger mixing at its upper surface and mixing by diffusive layering below. The subsurface entrained water has a high level of chlorophyll and dissolved oxygen and a species assemblage of diatoms distinct to the shelf species. The total annual export of biogenic carbon was estimated as $10^6 \text{ kg C yr}^{-1}$ and the total transport due to water entrainment as 0.15 Sv (Lillibridge et al., 1990).

Another mechanism of shelf–slope exchange in the MAB is related to warm-core rings (WCR) generated from Gulf Stream meanders (Evans et al., 1985; Garfield and Evans, 1987; Joyce et al., 1992). When the inshore edge of the ring comes in contact with the shelf–slope front, long filaments or streamers of shelf water are drawn offshore, wrapping anticyclonically around the WCR (Ramp, 1986). Estimates for the off-shelf transport associated with these streamers, based on in situ hydrographic data, current measurements with drifters and satellite SST images range between 0.001 and 0.2 Sv (Morgan and Bishop, 1977; Bisagni, 1983). These estimates show the variable nature of the inflow and outflow of shelf water in the MAB. Locally formed shelf break eddies also have a potential to enhance across-front diffusivity of heat and salt (Garvine et al., 1988). The shelf break eddies, with a wave length of 33 km, deform the front into ‘backward breaking’ waves. They cause plumes of lighter shelf water to protrude well over the outer slope contradicting the interpretation by Cresswell (1967) and Wright (1976) as detached parcels of shelf water.

SST images also show large outbreaks of shelf water over the slope region, not related to eddies or filaments (Garfield and Evans, 1987). These outbreaks are often found in the area around Northeast Channel and seem to originate from the Scotian shelf and the eastern end of Georges Bank. They appear as large tongues of shelf water protruding from the frontal area into the slope water. Smaller outbreaks retreat back into the shelf water, but larger ones seem to mix with the slope water. These outbreaks are observed 29% of the time and are not associated with Gulf Stream rings and eddies.

A potentially important reason for understanding the offshore extent of the shelf–slope front and the accompanied changes in shelf water volume in the MAB arises for its effect on the phytoplankton population and carbon and nitrogen budgets. In winter and early spring, the wedge of shelf water between the shelf break and the shelf–slope front may contain $>2 \mu \text{M NO}_3$, and phytoplankton blooms develop far seaward of the shelf break (Ryan et al., 1999). The primary production in these offshore blooms over the slope are initially fueled by nutrients transported from the shelf. Much of the carbon fixed in this wedge must sink through the shelf–slope front into slope water. Therefore, it is necessary to include all of the shelf water and not be limited to water lying over the continental shelf in the budgets (e.g., carbon and nitrogen) for the MAB.

The strong horizontal surface temperature (SST) gradients present in the southern MAB for most of the year (except during summer) are readily identified in satellite images. SST maps separated by only a few days portray motions and realignment of boundaries between different water masses in the MAB. In the southern MAB, the main boundaries separate shelf, slope and Gulf Stream waters. The shelf water–slope water boundary in particular is frequently found well offshore of the 100 m isobath (the conventional separation line
between “shelf” and “slope”). We call such large excursions of the front, overrunning of slope water by shelf water. The frequency and extent of such overrunning yield important clues to shelf circulation. They also provide information on shelf–slope exchange, fate of shelf water and on the shelf water budget.

In this study, we examine these overrunning episodes of the slope sea by shelf water in the MAB. The outline of the paper is as follows: Data and methods are described in Section 2. In Section 3.1, high-resolution satellite derived SST images are used to describe shelf water overrunning the slope sea in the southern MAB for two years, 1992 and 1993. Hydrographic cross-sections are then examined to relate surface features in the vertical dimension in Section 3.2. The combination of satellite and in situ data are then used to estimate the velocity, volume and transport of the overrunning water in Sections 3.3 and 3.4. Mechanisms affecting the export of this shelf water are examined in Section 3.5. Conclusions are given in Section 4.

2. Data and methods

2.1. In situ observations

Surface temperature, salinity and XBT profiles are routinely collected from the “ship of opportunity” R/V Oleander in a NY – Bermuda transect (Benway et al., 1993a,b). The data are obtained from National Marine Fisheries Service, NOAA for comparison with coincident satellite imagery.

2.2. Historical observations

The historical data set consists of cross-sections taken during the SEEP I and II and the National Marine Fisheries Service Marine Resources Monitoring, Assessment and Prediction (MARMAP) programs to the MAB (Biscaye et al., 1994; Mountain, 1991; Manning, 1992). In addition, a compilation of hydrographic transects of the MAB shelf–break front that describe the variability of the physical conditions in this region, and the shelf–slope front in particular, are also used in the analysis (Lyne and Csanady, 1984).

2.3. Wind observations

Hourly wind data, obtained from NDBC buoys located in the MAB (see Fig. 1) are used to analyze the shelf water response to wind stress. Alongshore and offshore stress components are computed by rotating the coordinate system 30° clockwise so that the alongshore direction was aligned with the large-scale bathymetric contours. Time series of wind stress are computed according to Large and Pond (1981). In all, data from five buoys are used to compare windstress in the northern and southern part of the MAB area. The coherence squared between the stresses at different locations is significant at the 95% confidence level at most frequencies. The phase is not significantly different from zero. Therefore, only data from buoy 44008 (for 1992) and 44014 (for 1993) are used to represent wind conditions for these two years in the MAB.

2.4. Estimation of alongshelf velocity

The velocity of shelf water overrunning the slope region is determined from sequential satellite SST imagery using the method described in Breaker et al. (1994). This method calculates the velocity of a particle, given its displacement between two successive images and the time interval between the two images assuming that the motion of the particle is linear. The above technique has been used extensively in meteorology for the last 25 years to estimate low-level winds from geostationary satellite data. It was also used for case studies on the Slope Sea and was found to be quite reliable. Cornillon et al. (1986) used the same technique to estimate the surface velocity of the slope waters and the translation velocity of Warm Core Rings.

In this study, SST images that are less than 24 h apart are selected to determine surface current vectors. The features used for tracking had to be defined by closed thermal contours; kinks and cusps are excluded. They are subjectively digitized and their center positions located. Small features are selected as they best represent the velocity of the surface slope water. Most of the features chosen are located in the slope region to determine the rate of advection of the shelf water that has overrun the Slope Sea.
The method assumes that sea surface temperature is a conservative passive tracer, neglecting changes due to surface heat exchange or diffusion. Propagating waves tend to interfere with the detection of advective motion in places where both are present. Furthermore, the motion of the particles are assumed rectilinear when in fact they could be rotational. The time separation between images is also an important criterion. When it is too long, features begin to lose their identity and it becomes difficult to assign a displacement. Svejkovsky (1988) found that intervals greater than 12 h but not more than 24 h between images produce the best results. Another problem is the lack of precise navigation, which can cause errors of 0.02–0.03 m s$^{-1}$. Procedures such as those developed by Breaker et al. (1994) typically reduce this error by a factor of 4. Landmarks are often useful in reducing these errors over coastal areas. In this study, all images are registered to a common coastline thereby reducing the error due to navigation. The uncertainty in the velocity estimate for feature tracking in this study for an average velocity of 0.2 m s$^{-1}$ is about ±0.04 m s$^{-1}$.

3. Results

3.1. Satellite Observations

A sequence of adequately clear images spanning the two years 1992 and 1993 are captured in eleven plates. The following is a running commentary of plate 1 and plate 2 (Fig. 2a–h), emphasizing their most notable features.

![Plate images](http://www.rsmas.miami.edu/personal/akumar/ajoyweb.pdf) See text for details. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
On 8 January 1992 (Fig. 2a) low sea surface temperature of about 4–8 °C is evident near the coast, progressively increasing to warmer temperatures of 10 and 12 °C in the slope region. The Gulf Stream is partly obscured by clouds. The 10° isotherm shows the shelf water-slope water surface front (“front” from here on) to be well within the 100 m isobath. The SST image also clearly shows shelf water entrainment to the Gulf Stream at Cape Hatteras. In the north, shelf water is seen, albeit partly covered by clouds, over the slope region. The SST image on 31 January (Fig. 2b) shows the front hugging the 100 m isobath from New Jersey to Virginia, where shelf water turns towards the Gulf Stream. This is presumably the undisturbed state of the front. In sharp contrast, the northeast corner of the image shows shelf water overrunning the Slope Sea to a distance of about 100 km off New Jersey. An anticyclonic eddy is also visible, just ahead of the shelf water in the slope region.

The next three images (Fig. 2c–e) display the movement of the boundary of the shelf water mass that overran the slope region. It has progressed southwestward by 50 km from 3 February to 10 February and by nearly 120 km by 17 February. The anticyclonic eddy has progressed similarly, projecting shelf water ahead on its journey southwestward. Eddy activity appears to mix some of the shelf water as seen by the light blue enclosures within the overrun shelf water. On 17 February, the anticyclonic eddy is connected to the Gulf Stream by a ‘streamer’ of shelf water. The anticyclonic eddy has almost disappeared by 21 February (Fig. 2f). Shelf Water off Delaware almost reaches the Gulf Stream. Most of the Slope Sea is covered by shelf water. Again, the Gulf Stream can be observed to entrain shelf water off Cape Hatteras.

On 28 February (Fig. 2g), part of the shelf water volume appears mixed with the slope water as seen by the light blue patches on the image. The large volume of shelf water has now progressed southwestward and has reached off Chesapeake Bay. Filaments from the Gulf Stream are connected to the shelf water at various points before being entrained by the Gulf Stream. Somewhat smaller but still substantial overrun of shelf water persists and is noticeable all along the front at the end of March (22 March, Fig. 2h).
3.2. In situ observations

The satellite derived data provide a synoptic coverage in both space and time. Hydrographic data, coincident with satellite data, are now examined to relate surface features with hydrographic cross-sections in the vertical dimension.

The upper panel of Figs. 3 and 4 shows shipboard surface salinity and temperature along with temperature extracted from satellite infrared images for approximately the period of each cruise. At least three distinct water masses can be related to surface SST images in our discussion above, shelf water, slope water and Gulf Stream water. In particular, the fronts show the same structures, but the satellite profiles show greater variation and underestimate the surface temperature, in some cases up to 5°C. The difference between satellite and in situ estimates is due to two reasons. First, the satellite measurements lag or precede the in situ observations by a few days. Clouds in satellite measurements that appear along the transect or near it can cause substantial underestimation. Second, in situ surface temperatures are sampled 10–20 km apart (XBT spacing), whereas the satellite show finer-scale variations with a pixel resolution of about 1.1 km.

The shelf–slope front is represented by the 10°C isotherm in SST images (Fig. 2) and can be identified by the thermal gradient between shelf (7°C) and slope (12°C) waters in in situ data during winter and early spring. The shelf–slope front signature can also be observed in salinity profiles by the gradient between shelf water (33) and slope water (35.3). This boundary between shelf and slope waters will be used later to determine the offshore limit of shelf waters in the MAB.

The lower panels of Figs. 3 and 4 show vertical cross-sections from XBT data along the same transect. The Oleander section of February, 1992 (Fig. 3) shows that the front extends into the slope regime about 150 km
from the shelf break. This can also be seen on the SST image of February 3rd (Fig. 2c). The base of the front intersects the 80 m isobath. In addition, the front encloses shelf water to a depth of ~60 m offshore of the shelf break. The shelf water properties are vertically uniform down to 60 m. Lenses of 13 °C water are observed from the surface to a depth of 200 m and intrude onshore over the continental slope. The Oleander data, therefore, illustrates that the shelf water mass observed on the SST image of February 3rd reaches depths of up to 60 m offshore of the shelf break.

The transect in March, 1992 (Fig. 4) shows the shelf–slope front ~50 km offshore of the shelf break, extending to a depth of ~50 m. The XBT contours show steeply sloped isotherms that suggest strong southwestward flow. In the slope region, the 13° thermostad reaches to a depth of 150 m. Once again the shelf water is vertically well mixed at this period of the year mainly due to strong winds and weak buoyancy input.

The hydrographic cross-sections from Oleander combined with earlier evidence (e.g., historical data) illustrate that overrunning of the slope region by shelf water observed in SST images reaches depths of 40–60 m. The overrun shelf water, bounded by the shelf break and the shelf–slope front offshore enclose a large portion of the total shelf water volume in the MAB.

3.3. Estimation of overrunning episodes

From the satellite imagery comprising 18 usable images for the two years, the shelf water overrunning and advection along the Slope Sea are seen to occur in pulses. When one pulse of shelf water leaves the region off New Jersey it is tracked in a sequence of successive SST images until it entrains eastward between Chesapeake Bay and Cape Hatteras. The time interval from its inception to when it turns eastward is defined here as an episode.

The first episode of overrunning of shelf water during 1992 started around January 31. The feature tracking showed persistent velocities of 0.2 m s⁻¹ just off New Jersey in the southwestward direction. The shelf water then slows down to velocities of 0.05 m s⁻¹ probably due to the motion of an anticyclonic eddy. The anticyclonic eddy entrains and wraps cold shelf water as it advects southwestward. This first overrunning episode
lasted about 34 days. Velocities of 0.2 m s\(^{-1}\) are also observed near the southernmost tip of the Slope Sea, probably due to the influence of the Gulf Stream.

The next episode of overrunning originates off New Jersey on February 17th (Fig. 2e). The influence of the second episode on the shelf water mass can be recognized in successive images until the 28th of March especially on the 15th and 22nd of March, lasting for about 39 days. The shelf water advection is again southwestward along the upper slope region and often reaches velocities of 0.2 m s\(^{-1}\), especially off New Jersey.

The third episode begins around the first of April and by the 15th of April it has reached off the coast of Delaware. Unlike the earlier episodes, the shelf water mass entrains much earlier into the western edge of the Gulf Stream and does not reach Cape Hatteras. The shelf water again advects along the upper slope region with velocities ranging from 0.05 m s\(^{-1}\) to nearly 0.2 m s\(^{-1}\).

The first episode in 1993 started around January 24th. The shelf water had a maximum velocity of \(~0.20\) m s\(^{-1}\) near New Jersey and advected southwestward. The progression could be followed to March 1st, before it entrained in the vicinity of Cape Hatteras, lasting \(~35\) days.

The next episode started on March 1st before the end of the first episode and is clearly observable on the image of 12th March. The entrainment of this episode occurs much earlier off Chesapeake Bay and does not advect to Cape Hatteras. This shorter episode lasted around 22 days with a maximum velocity of \(~0.20\) m s\(^{-1}\) occurring off New Jersey.

The third episode began around March 22nd and could be followed up to April 8th. The image on the 8th shows another episode off New Jersey. The third overrunning episode lasted until April 23rd for about 31 days. The shelf water in the slope region off New Jersey again showed a persistent southwestward velocity of \(~0.20\) m s\(^{-1}\).

The two year analysis showed that (three to four) episodes can be identified during a period of approximately three months during winter and early spring. Each episode lasted about 30 days. The episodes of shelf water overrunning the Slope Sea originate off New Jersey (at the New York Bight apex) around the end of January and advect southwestward along the upper slope region. The maximum velocities (\(~0.20\) m s\(^{-1}\)) were recorded off New Jersey and decreased slightly as the event advected towards Cape Hatteras. The entrainment of shelf water to the Gulf Stream or the Slope Sea takes place anywhere between Cape Hatteras and off Chesapeake Bay and advected eastward along the western side of the Gulf Stream.

The shelf water traveled a distance of \(~500\) km in the slope region from New Jersey to near Cape Hatteras in about 30 days. This gives a mean velocity of \(~0.19\) m s\(^{-1}\) which is consistent with the mean velocity estimated from feature tracking.

The velocity estimate from feature tracking is almost twice the mean velocity (\(~0.10\) m s\(^{-1}\)) observed in the outer shelf region of the MAB (Beardsley and Flagg, 1976). This suggests that the advection of shelf water over the upper slope region may be linked with the circulation of the Slope Sea. The mean southwestward velocity of the upper slope current off New Jersey is about 0.15 m s\(^{-1}\) (Csanady and Hamilton, 1988). The relative velocity by geostrophic calculation of shelf water layer is about 0.1 m s\(^{-1}\). This gives a good reason to believe that the overrun shelf water motion over the upper slope region, bounded by the front offshore and to a depth of \(~50\) m, is carried southward by the upper slope current.

3.4. Volume and transport of overrun shelf water

To estimate the volume of shelf water overrunning in the southern MAB, the offshore extent of shelf water is determined from the position of the shelf–slope front in each SST image. Previous studies have shown that the location of the shelf–slope front at the surface as determined by satellite imagery could account for about 50% of the variance in the shelf water volume in the MAB (Mountain, 1991). In this study, the area from the shelf break to the 10 °C contour line is used to estimate the volume of shelf water that overruns the slope region in the southern MAB. The area bounded by the shelf break and the 10 °C contour is obtained by digitizing the coordinates of each pixel in the image along this line and by integrating up to 40°N.

The depth of the shelf water in the Slope Sea varies generally from 40 to 60 m. The Oleander XBT data showed the overhang between 50 and 60 m deep. Sections from SEEP programs taken off New Jersey and off Delaware also showed the depth of the overhang to vary between 40 and 60 m. Historical transects and cross-sections taken from “A compilation and description of hydrographic transects of the Mid-Atlantic Bight
Shelf–Break Front” atlas (Lyne and Csanady, 1984) which span more or less the whole of the MAB have also confirmed the depth of the overhang to vary within 40–60 m. Mountain (1991) used an average depth of 40 m in his calculation of the volume of shelf water from cruises between 1977 and 1987.

An uncertainty in the volume estimate can be induced by the constraint in assuming a constant depth of \( \sim 40 \) m and the error in determining the area of the overrun shelf water. An error of about 10 pixels will give an uncertainty in the area of about \( \pm 120 \) km\(^2\). The volume uncertainty for depths of 40, 50 and 60 m is about \( \pm 4.8, 6 \) and 7.2 km\(^3\).

The volumes of shelf water for the three episodes during 1992 are shown in Fig. 5. The first episode of overrunning reached a maximum of about 1000 km\(^3\). The volume of shelf water for the second episode reached a maximum of about 2100 km\(^3\) around day 80, after which it decreased steadily until about day 100. The second episode thus lasted longer and contained more volume than the first episode. A feature in the third episode is the patchiness in the shelf water mass that has overrun the slope region. This is noticeable by the presence of slope water in the region off New Jersey and below it. This large estimate is partly due to the stagnant shelf water mass in the slope region as seen in the SST images.

The volumes of the overrun shelf water mass and the episodes during 1993 are shown in Fig. 6. Although the overrunning started slightly earlier, the volume was considerably less than in 1992. The maximum volume (2000 km\(^3\)) again occurred at the end of March which may be due to the accumulation of shelf water in different episodes as in 1992. Thereafter, the shelf water volume decreased steadily to about 1000 km\(^3\) in late April.

The mean volume of the overrun shelf water (between the shelf break and the location of the shelf–slope front) is about 1600 \( \pm 55 \) km\(^3\) for 1992 and 900 \( \pm 55 \) km\(^3\) for 1993. The mean shelf water volume in the MAB is about 3600 km\(^3\). This shows that about 25–45% of the volume of shelf water in the MAB lies offshore of the shelf break during these episodes. Wright and Parker (1976) observed that about 50% of the volume of shelf water in the MAB lie in the outer wedge between the shelf break and the shelf–slope front. The large variation in volume between 1992 and 1993 (\( \sim 70\% \)) may be associated with increase in the flow into the Gulf of Maine and later into the MAB. Mountain (1991), in his 10 year analysis of the volume of shelf water in the MAB has observed interannual changes of 60–90% from the mean volume in the MAB.

![Shelfwater Volume from Satellite Data for Each Overrunning Episode (1992)](image)

Fig. 5. The shelf water volume between the 100 m isobath and the satellite determined position of the shelf–slope front for the three overrunning episodes during 1992. The three episodes are shown using different symbols.
The transport of the overrun shelf water just off New Jersey can be estimated from the Oleander cross-sections and the velocity determined from feature tracking. On the 8th of February, 1992, the shelf water extends ~100 km offshore. Taking an average depth of about 40 m and ~0.2 m s\(^{-1}\) as the cross-sectional velocity, the transport is ~0.8 Sv. Taking a depth of 60 m, the transport would be 1.3 Sv. Similarly, on the 11th of March, the shelf water extends about 60 km offshore and the transport associated with it is ~0.5 Sv and 0.7 Sv for \(h = 60\) m. During 1993, the transport estimate from 6th February and 8th April is again ~0.8 Sv. This gives an average of ~0.7 Sv for one third of a year for two episodic events or the episodic annual average may be about 0.12 Sv. The high transport estimate suggests that the transport of the overrun shelf water is large and cannot be ignored. The uncertainty associated with this transport is related to the velocity measurements and the depth of the overrun shelf water as discussed before. For a maximum offshore distance of ~100 km the uncertainty in the transport estimate is ±0.64 Sv. Our procedure clearly gives only rough approximations of shelf water transport but indicates how shelf water transport varies over the sloperegion in the MAB.

3.5. Forcing mechanisms

In this section, we are concerned with identifying forcing mechanisms responsible for overrunning events such as wind stress, freshwater discharge, pressure gradient effects and baroclinic or barotropic instability.

3.5.1. Effect of windstress

To illustrate the variability of the magnitude of the stress for 1992 and 1993, the alongshore and offshore components are shown in Figs. 7 and 8. Maximum stresses occur during mid January and early February. In addition, the winter stress tends to be directed offshelf with brief wind reversals. The large stress values also arise principally because of frequent strong winds rather than because of the tendency of the winds to blow more frequently in a given direction.

Spectral analysis (not shown) for 1992 reveal that energetic events for the offshore windstress occur at 2, 5 and 17 days period with the most energetic period between 4 and 5 days. In the alongshore direction, the
The dominant period is 4 days with two smaller peaks at 3 and 7 days. The spectral analysis for 1993 offshore wind-stress shows variability of 3 and 10 days with most of the energy concentrated at 10 days. In the alongshore direction the variability is at 3, 4 and 17 days with the most energetic period being 4 days. However, the striking feature of the wind stress in the MAB is its variability and there are periods of strong wind reversals during the entire four month periods with alongshore winds to the north of greater than 0.1 Pa for both the years.

The work done by the alongshore component of wind stress on the overrunning of shelf water in the MAB can be examined in terms of the time integral of the wind stresses \( \frac{1}{f} \int \tau_x dt \) and the Bakun index. \( \int \tau_x dt \) is the cumulative sum of the magnitude of hourly wind stress vectors, where \( \tau_x \) is the kinematic wind stress \( \frac{w}{f} \) in the alongshore direction. The Bakun index or the upwelling index is the off-shore component of Ekman transport derived from surface winds (Bakun, 1976). The Ekman transport is calculated from \( \frac{1}{f} D \int \tau_x dt \) where \( f \) is the Coriolis parameter. The offshore displacement can be expressed as \( \frac{1}{f} D \int \tau_x dt \) where \( D \) is the Ekman depth. A comparison of time integral of the wind stresses \( \int \tau_x dt \) and SST images showed that cross-shelf motions of the front can be wind driven (Fig. 9). A northward wind starting on day 5 and another on day 23, moved shelf water and the shelf–slope front upto the slope region as seen in Fig. 2a and b. The most intense ocean response to strong northward windstress from day 40 to around day 61, produced a displacement of ~4 km for a depth

Fig. 7. Time series of along- and across-shore wind stress from buoy 44008 for the year 1992.
of 50 m and \( \sim 20 \) km if only 10 m of the water column is taken into account. The displacement determined from satellite imagery (Fig. 2e–g) is more than 50 km. The intense storm from day 70 to around day 80 again moved shelf water offshore all along the shelf–slope front (Fig. 2h) with an offshore displacement in the range of 6–30 km for depths of 50 m and 10 m, respectively. The image on 22 March show an offshore displacement of more than 50 km. The less intense northward winds from day 90 onwards also helped in driving shelf water into the slope region. The most energetic northward wind events for 1993 (Fig. 10) occurred from day 60 to day 75 and from day 97 onwards. The wind induced displacement per unit length of coastline was \( \sim 8 \) km and 40 km for depths ranging from 50 to 10 m, respectively. These observations illustrate that although northward winds move shelf water offshore, the displacements measured from satellite imagery are greater than those determined from wind stress only.

Another observation that can be made from satellite imagery and the wind data are that the ocean response to periods of strong southward wind stress does not appear to always displace the shelf–slope front significantly onshore. The shelf water that was transported offshore to the slope region moves southwestward and entrains off Cape Hatteras. For example, the strong southward winds in 1992 during days 32–42 did not move the shelf–slope front or for that matter, shelf water mass, significantly onshore (Fig. 2c and d). During SEEP I and II, the sequence of northward and southward events on the shelf produced large fluctuations

![Wind stress from buoy 44014-1993](image)

Fig. 8. Time series of along- and across-shore wind stress from buoy 44014 for the year 1993.
Fig. 9. A comparison of Bakun Index with frontal movement are shown for 1992. Periods of strong northward windstress are shown as horizontal lines and the satellite determined displacement of the front are shown as black dots in the figure.

Fig. 10. A comparison of Bakun Index with frontal movement are shown for 1993. Periods of strong northward windstress are shown as horizontal lines and the satellite determined displacement of the front are shown as black dots in the figure.
Fig. 11. The river discharge from the St. Lawrence River during 1991 and 1992 along with the mean. The large discharge during 1991 contributed to the overrunning of shelf water off NJ.

Fig. 12. The river discharge from the Chesapeake Bay during the three years; 1991, 1992 and 1993 along with the mean. The large discharge during 1993 contributed to the overrunning events in the southern MAB.
of the shelf–slope front and the front rapidly returned to its equilibrium position. But this was not always the case at the shelfbreak (Houghton et al., 1994).

3.5.2. Effect of freshwater discharge

The volume changes of shelf water in the MAB also are influenced by changes in the major inflows to the Gulf of Maine (Mountain, 1991). Factors controlling the amount of shelf water available for flow into the Gulf of Maine during early winter–spring include the St. Lawrence River discharge during the previous spring. An increase in the St. Lawrence discharge, results in an excess quantity of shelf water (after mixing with slope water in the Gulf of Maine) flowing into the MAB.

To examine whether freshwater discharge from the St. Lawrence River produced the flooding of shelf water observed off New Jersey, the monthly-averaged St. Lawrence River discharge data measured at “LaSalle” (Station No. 020A016) for 1991 and 1992 and also multi-year long-term monthly averages (computed from the beginning of 1955 through the end of 1992) are shown in Fig. 11 (Bisagni et al., 1996; Bisagni and Smith, 1998). The cumulative freshwater discharge from the St. Lawrence increased by 15.1 km$^3$ during the first five months of 1991 relative to the long term mean (115.17 km$^3$). The freshwater mixes with more saline slope water in the Gulf of Maine and results in an $\sim$225 km$^3$ increase in shelf water entering the MAB by

Fig. 13. A 4-month-long time series (for 1992 and 1993) of the surface position of the Gulf Stream and shelf water on a line extending directly offshore from the shelf break (100 m isobath). The Gulf Stream position varies from $-20$ to 60 km, whereas the shelf water distances vary from 0 to 140 km from the shelfbreak. $r = -0.806$. This relationship demonstrates that a change in the Gulf Stream position has a blocking effect on the shelf circulation causing the overrunning events.
late January 1992. The time lag between the St. Lawrence discharge and the inflow into the MAB of about nine months (Bisagni et al., 1996), is sufficient for the excess outflow to reach the vicinity off New Jersey. Therefore, the shelf water flooding the Slope Sea off New Jersey during late January 1992 appears to be related to excess discharge from the St. Lawrence in the previous months. To examine if the outflow from the Chesapeake Bay can cause any significant contribution to the overrunning of shelf water, the monthly averaged discharge from all the rivers that flow into the Chesapeake Bay along with a long–term mean (computed from the beginning of 1950 to the end of 1988) are shown in Fig. 12. During spring of 1993 an anomalous large freshwater discharge from the Chesapeake Bay of about 10,000 m$^3$ (an increase of about 6000 m$^3$ from the mean) is observed. Since the discharge from the Chesapeake Bay occurs further south it cannot have contributed to the overrunning of shelf water off New Jersey. However, the changes in the outflow from the Bay can account for the increased volume of shelf water further south in the MAB.

3.5.3. Pressure gradient effect

The alongshelf pressure gradient in the MAB is an important force in the shelf circulation (Csanady, 1976). The mechanism that creates the pressure gradient originates in the open ocean. It is interesting to examine how remote offshore forcing, such as the Gulf Stream can alter the pressure gradient and what effect it can have on the shelf water circulation.

The instantaneous path of the Gulf Stream in this region may be complex. Several mechanisms cause the path to move laterally and can induce shifts of 100 km or more in the Stream’s path in a few days. Such shifts in the path of the Gulf Stream have shown to discharge fluid from the Stream into the shelf and slope region in the MAB (Churchill and Cornillon, 1991b). Bane et al. (1990), have also shown that such onshore movements of the Stream strongly increase the mean monthly shelfbreak southwestward currents in the MAB.

Satellite-derived information on the position of the Gulf Stream’s shoreward surface thermal front (Fig. 1) and the offshore limits of shelf water in the slope region off New Jersey suggest a relationship between the overrunning events and the Stream’s position. Fig. 13 shows the 4-month-long time series for 1992 and 1993 of the surface position of the Gulf Stream and shelf water on a line extending directly offshore from the shelf break (100 m isobath). Both positions are shown in Fig. 1. Since the shelf water overrunning was maximum off New Jersey, this location was used for the offshore position of shelf water. Negative distances for the Gulf Stream position indicate water onshore of the shelfbreak. Note that the Gulf Stream position varies from $-20$ to 60 km, whereas the shelf water distances vary from 0 to 140 km from the shelfbreak.

The agreement between the two time series appears good. These data suggest that the shelf water overrunning the Slope Sea off New Jersey was extensive when the Gulf Stream was nearer the shelfbreak in the vicinity of Cape Hatteras and decreased as the Stream veered offshore. This relationship demonstrates that a change in the Gulf Stream position has a blocking effect on the shelf circulation causing shelf water to overrun the shelf break and move offshore into the slope region.

In summary, the shelf water overrunning the slope region of the southern MAB is a complex interaction of various forcings which include wind stress, freshwater discharge and the remote offshore forcing from the Gulf Stream.

4. Conclusions

This study examined the seasonal variation of shelf and slope waters in the MAB using two years of high-resolution sea surface temperature data derived from NOAA satellites. Additionally, Oleander transects that coincided with the satellite images combined with earlier hydrographic data were used in the analysis of the images.

The two year (1992 and 1993) SST analysis revealed extensive overhang of shelf water into the slope region. These protrusions of shelf water occurred episodically and coherently along shelf distances of several 100 km. The satellite images combined with in situ data further illustrated that the overrun shelf water, bounded by the shelf–slope front offshore and beneath, enclose a large volume of shelf water. The shelf water was advected southwestward in several episodes and caused extensive overrunning of the Slope Sea. The large shelf water volume produced by overrunning events was transported southwestward along the upper slope and later accounted for the entrainment of shelf water by the eastward flowing Gulf Stream in the vicinity
of Cape Hatteras. The focus of this work was on estimation of the velocity, volume and transport of the overrun shelf water from satellite images and in situ data. The velocity of shelf water overrunning the slope region was determined from sequential satellite imagery using the feature tracking method. The satellite feature tracking showed a maximum velocity (≈0.20 m s\(^{-1}\)) off New Jersey which decreased slightly as the episodic events advected towards Cape Hatteras. Three to four episodes could be identified during a period of approximately three months from winter to spring. The episodes of shelf water overrunning the Slope Sea originated off New Jersey at the New York bight apex around the end of January and was advected southwestward along the upper slope region. Each episode lasted about 30 days. The high rate of advection of the overrun shelf water over the upper slope region suggests that the shelf water, bounded by the front offshore and to a depth of ≈50 m, could be carried southwestward with the help of the upper slope current along the western flank of the Slope Sea.

The volume of shelf water that overruns the slope region showed a general increase from late January to a maximum of ≈2000 km\(^3\) during 1992. During 1993, although the overrunning started earlier, the volume was considerably less than in 1992. The maximum volume (≈2000 km\(^3\)) again occurred during the end of March which may be due to the accumulation of shelf water in different episodes as in 1992. The shelf water volume decreased until the end of April as the shelf–slope front retreated back to its normal position over the shelf break. The mean volume of the overrun shelf water was about 1600 km\(^3\) for 1992 and 900 km\(^3\) for 1993. This volume comprised between 25% and 45% of the total volume of the MAB and showed that an important component of the shelf water volume in the MAB actually lies offshore of the shelf break.

An estimate of the transport yielded an annual average of about 0.12 Sv. The shelf water was transported along and across the Slope Sea and its fate was largely influenced by the gyre circulation in the vicinity of Cape Hatteras. The SST images showed that the overrun shelf water was entrained intermittently into the Slope Sea at various locations between 38°N and Cape Hatteras and eventually drained into the eastward flowing branch of the Gulf Stream. Forcing of the shelf water, causing it to overrun the slope region were also examined. Northward winds appeared to force the shelf water offshore to distances of ≈10 km to 40 km. However, the satellite determined offshore extent of the overrun shelf water was larger than that produced by wind events. This showed that wind forcing of shelf water was not solely responsible for large overrunning events. The fresh water discharge from the St. Lawrence was linked to the inflow of shelf water into the MAB. Interannual variations of St. Lawrence discharge related to the shelf water flooding of the Slope Sea off New Jersey. Another likely mechanism that could contribute to the overrunning events is the lateral displacement of the Gulf Stream in the vicinity of Cape Hatteras. The fluid discharged by the Gulf Stream produces a blocking effect on the shelf circulation. This resulted in shelf water overrunning the slope region especially off New Jersey. In conclusion, the shelf water overrunning the slope region in the southern MAB is a complex interaction of various forces including wind stress, fresh water discharge and the blocking effect on shelf circulation due to the onshore veering of the Gulf Stream.

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