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Interannual variations in upper-ocean transport by the Gulf Stream and adjacent waters between New Jersey and Bermuda

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ABSTRACT

Since the fall of 1992, an acoustic Doppler current profiler mounted on a freighter, the CMV Oleander, has been measuring upper-ocean currents between New Jersey and Bermuda on a weekly basis. The extensive database that results from the frequent, systematic, and sustained sampling enables the exploration of a number of questions regarding currents in the northwest Atlantic. This paper reports on interannual variations in transport in the Gulf Stream and adjacent waters. The repeat sampling greatly increases the ability to discern even rather subtle variations in near-surface transport and explore their possible causes.

The transect is divided into three subregions: the Gulf Stream is defined by a high velocity core with an instantaneous width set by where the downstream component of velocity changes sign; the Slope Sea exists between the Gulf Stream and the continental shelfbreak; the Sargasso Sea lies between the Gulf Stream and Bermuda. These three regions exhibit quite different signatures of variability. Over the eleven years of operation to date annual averages of Gulf Stream transport have a standard deviation of 6% but a 23% peak-to-peak range. No discernable trend in transport is evident in the eleven-year record. The westward transport in the Slope and Sargasso seas can both vary by a factor two in magnitude but they have quite different temporal characteristics: the Slope Sea transport changes take place gradually whereas the Sargasso Sea exhibits much larger variations on shorter time scales. It is conjectured that the Slope Sea time scales are set by high-latitude buoyancy-related forcing, whereas the Sargasso Sea and Gulf Stream variability reflects tropical and subtropical mechanical forcing.

The lateral position of the Gulf Stream exhibits a correlated behavior with westward transport in the Slope Sea. When Slope Sea transport increases, the Gulf Stream shifts to the south with a concomitant hint of increased Gulf Stream transport. The southward shift of the Gulf Stream may be part of a dynamical response to this increased circulation in the Slope Sea since the Slope Sea flow is blocked in the west by the Gulf Stream at Cape Hatteras suggesting that the path of the Gulf Stream is governed more by thermohaline- than wind-driven forcing. The fast time scales of transport in the stream, on the other hand, point to wind-driven forcing from the tropics and subtropics. Thus Gulf Stream position and transport would appear to be driven by quite different physical processes.

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1. Introduction

The advent of shipboard acoustic Doppler current profiler technology has ushered in a new era of ocean observation. These instruments, mounted on ships underway, have allowed oceanographers to measure currents with extraordinary resolution in both the horizontal and in the vertical, and without any assumptions about geostrophy or reference levels. Further, when the global positioning system (GPS) and somewhat later GPS-based compasses became available in the 1990s, the vessel's movement could be subtracted from the measured velocity with such accuracy that fluid velocity relative to the bottom could be estimated to within a few cm s^{-1}. This extraordinary measurement accuracy and spatial resolution has opened up a new approach to the study of ocean currents.

Since 1992 our group has been measuring upper-ocean currents from a container vessel—the CMV Oleander—that operates between New Jersey (NJ) and Bermuda. The instrument, an acoustic Doppler current profiler (ADCP), can resolve the horizontal structure of ocean currents in considerable detail, and by repeating these sections over and over again one can obtain exceptional insight into the spatial-temporal variability of currents in the northwest Atlantic. A major objective of the Oleander program is to track the variability of the ocean on time scales of years to decades. The program has now entered its 13th year of operation.

This program was initiated to take a closer and more quantitative look at the velocity field of the northwest Atlantic and the Gulf Stream in particular. Numerous hydrographic sections across the Gulf Stream had revealed a general stability of transport on the one hand, but considerable scatter between estimates on the other. It was difficult to sort out the temporal content of these observations with such sparse sampling. Indeed, it was possible to conjecture significant low-frequency variations in transport in the absence of meaningful estimates of measurement uncertainty. It became increasingly clear that a new approach was needed, one that would allow for greater sampling frequency and hence averaging for a given period of time. This paper examines the low frequency variability of upper-ocean transports in the northwest Atlantic along the Oleander transect with particular focus on the Gulf Stream.

2. Background

The long-standing definition of transport in the Gulf Stream was developed by Iselin (1936) where he computed baroclinic transport relative to a level of no motion at 2000 m. The horizontal limits of integration across the current were effectively set by where the pycnocline leveled out to either side of the current. Starting in 1939 Iselin organized a program of monthly crossings of the Gulf Stream, but it was ended prematurely by the onset of the 2nd world war (Iselin, 1940). Starting with the famous Gulf Stream '60 (Fuglister, 1963) project, numerous sections have since been taken across the Gulf Stream. Worthington in his 1976 monograph concluded from 32 sections that the baroclinic transport (0–2000 m) east of Cape Hatteras ranged between 85 Sv (1 Sv $= 10^6$ m^3 s^{-1}) in April and about 70 Sv in December.

A different approach for estimating Gulf Stream transport took place between 1980 and 1983 (Halkin and Rossby, 1985). Every two months a set of vertical profiles of velocity from the surface to the bottom across the Gulf Stream was obtained. These direct measurements of velocity were free of any assumptions about geostrophy and allowed a closer look at the dynamical structure of the current. From these sections a mean transport of 87.8 Sv was obtained, of which 73.4 Sv would remain as the baroclinic component if the observed velocity at 2000 m depth were removed from the profile. But even these directly measured transports were subject to a 20% range in scatter making it difficult to estimate the annual cycle with any accuracy not to mention any change on longer time scales.

Some twenty years subsequent to the Worthington study, Sato and Rossby (1995) reanalyzed 130 carefully screened hydrographic sections from between 1932 and 1988. They reported a 0–2000 dbar mean transport of 71 Sv with an 8 Sv annual range that peaks in summer. But more relevant to this study (Sato and Rossby, 1995) was one of very few attempts at the time to estimate low-frequency variations in transport. While they suggested the possibility of interpentadal variations, examination of their Figure 10 indicates that even with 130 sections at their disposal, little could be said about interpentadal and longer variations due to the large scatter between individual estimates. The average transports from the three best pentads, 1935–1939, 1980–1984 and 1985–1989, lie within 4 Sv of each other. No significant secular trend between the 1930s and the 1980s, the two best-sampled decades, could be detected. In a very different approach Greatbatch *et al.* (1991) used the mean density fields from the two different pentads (1955–1959 and 1970–1974 reported by Levitus, 1989) to estimate the corresponding circulation patterns. They estimated a 30 Sv decrease in Gulf Stream transport, but the spatial resolution of their fields did not allow them to distinguish between the Gulf Stream itself and variations taking place within the subtropical gyre. Both studies had to cope with serious data limitation issues; in the first case infrequent sampling, and in the second case low spatial resolution. A different approach was needed.

3. Oleander ADCP operation

The CMV Oleander, a container vessel owned and operated by the Bermuda Container Line (BCL), provides year-round weekly freight service between Port Elizabeth, NJ and Hamilton, Bermuda. Knowing that the BCL was building a new vessel, we inquired about the possibility of mounting an ADCP in the hull so that currents could be measured along the vessel's path. The company was very receptive to the idea and before the ship entered service a cofferdam and other hardware had been incorporated into the vessel. At the vessel's first dry docking in fall 1991 the housing for the ADCP (known as the seachest) was installed. The ADCP was installed in summer 1992 and data have been collected ever since (Rossby and Gottlieb, 1998; Flagg *et al.,* 1998). It took some time to debug the system due to the very special conditions of working on a commercial vessel, but the data recovery gradually increased and since 1994 when the transducer was reoriented and a

GPS-heading system added to provide continuous calibration of the shipboard gyrocompass, the data flow and quality have improved greatly.

The Oleander (as do all modern container vessels) has a rectangular cross-section. This makes it very easy to install the seachest. On the other hand, this type of hull shape means that the vessel draws relatively little water, about 4 m in the case of the Oleander. She draws a bit more when fully loaded on the outbound leg and about 0.5 m less on the return trip to NJ. The shallower draft also means a higher likelihood of bubble entrainment on northbound transits. As a result, data return and quality depend to some extent upon the vessel's load factor. Weather conditions matter too; in heavy seas bubble drawdown becomes a serious data-limiting factor.

The Oleander's ADCP is an RD Instrument 150 kHz narrowband system. The Doppler measurement depends upon the acoustic backscatter of its transmitted signal by planktonic material in the water column such that the higher the biomass at depth the deeper one can profile. In the Slope Sea with its high biomass, the ADCP reaches to 250 m or more whereas in the Sargasso Sea with its lower biomass a typical depth limit of 150 m is reached. The profiles have a vertical resolution (or bin size) of 8 meters in the open ocean. The greatest depths are typically reached at the nutrient-rich Gulf Stream north-wall. With a downtime of the ADCP system of less than 10%, we typically obtain 40 or so complete transits between NJ and Bermuda per year. The reader is referred to the paper by Flagg *et al.* (1998) for a thorough treatment of the shipboard ADCP system and its operation. The paper by Rossby and Zhang (2001) gives a detailed discussion on the velocity and vorticity structure of the Gulf Stream.

4. ADCP data

The objective of this paper is to examine the long-term variability of the near-surface circulation in the northwest Atlantic between the continental shelfbreak and Bermuda. The database consists of a growing ensemble of ADCP sections. Figure 1 shows a typical transit from NJ to Bermuda with velocity vectors at 52 m depth superimposed on a sea surface temperature (SST) image from the same time. In addition to the Gulf Stream, a warm and a cold core ring can be seen in both velocity and SST. The velocity maximum of the current has remarkably tight limits at $2.07 \pm 0.24\ \mathrm{m\ s^{-1}}$ regardless of position, direction of flow, or time of year (Rossby and Zhang, 2001). The geographical average of all sections, shown in Figure 2, reveals clearly the Gulf Stream flowing off to the ENE and the return flows both north and south of the current. Due to the meandering, the peak speed is less and the current appears wider than it would in any single crossing. Note that the variance ellipses toe in toward the center of the current. This reflects, we think, decreasing width of the meander envelope in the downstream direction. That the Gulf Stream appears so sharply in this figure reflects the rather felicitous fact that the Oleander crosses the current close to, and just west of, a minimum in the meander envelope (Cornillon, 1986; Song *et al.,* 1995). To the south and north of the current except near the continental and

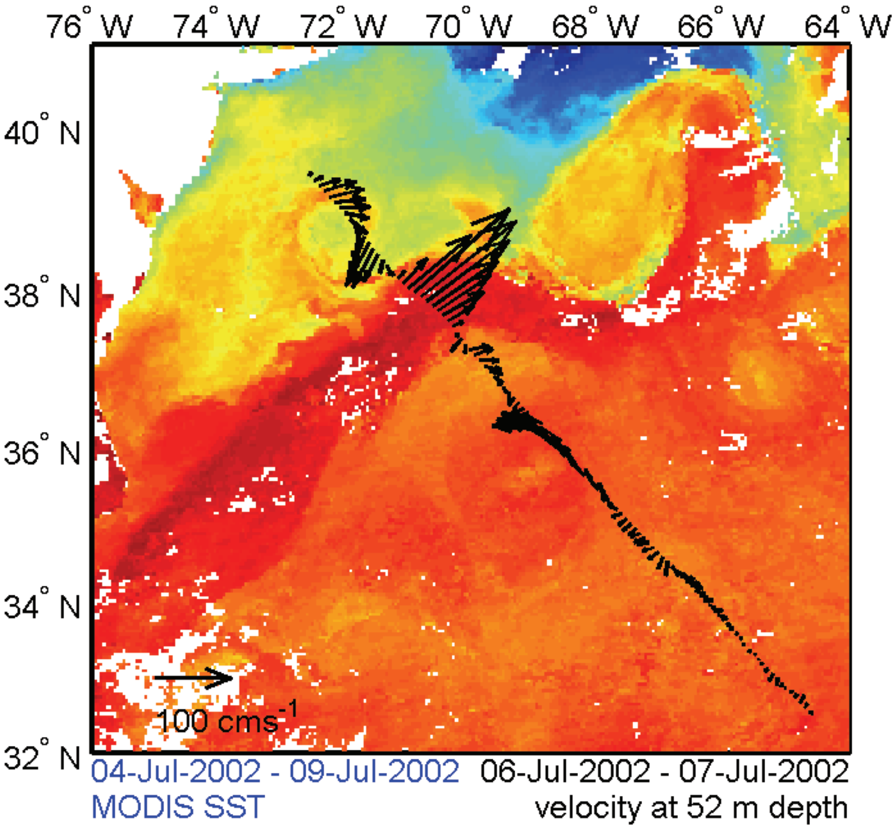

Figure 1. A typical section of ADCP velocity vectors at 52 m depth from the Oleander superimposed on a composite AVHRR sea-surface temperature image from July 2002. This period was chosen to illustrate not only the Gulf Stream but also the presence of warm and cold core rings in both data sets. White areas indicate land or clouds.

Bermuda slopes, the variance ellipses exhibit less polarization reflecting a more isotropic eddy field.

An effective way to gain insight into the low-frequency meandering of the current can be obtained from a time-distance or Hovmöller diagram of velocity in the mean downstream direction (67°T), Figure 3. This representation clearly reveals the Gulf Stream as a stiff structure gradually shifting north and south with time. (In preparing this figure the sections were first grouped into three-month averages. These were then further smoothed with 1/4, 1/2, 1/4 weights.) The width of the current, as indicated by the heavy black lines to either side, remains relatively invariant, particularly on the cyclonic side with its sharper velocity gradient. The stream can be seen shifting south during the 1996–1998 years followed by a large northward shift until the middle of year 2000 where the stream reached its most northerly extent in over 20 years based on expendable bathythermograph (XBT) data from

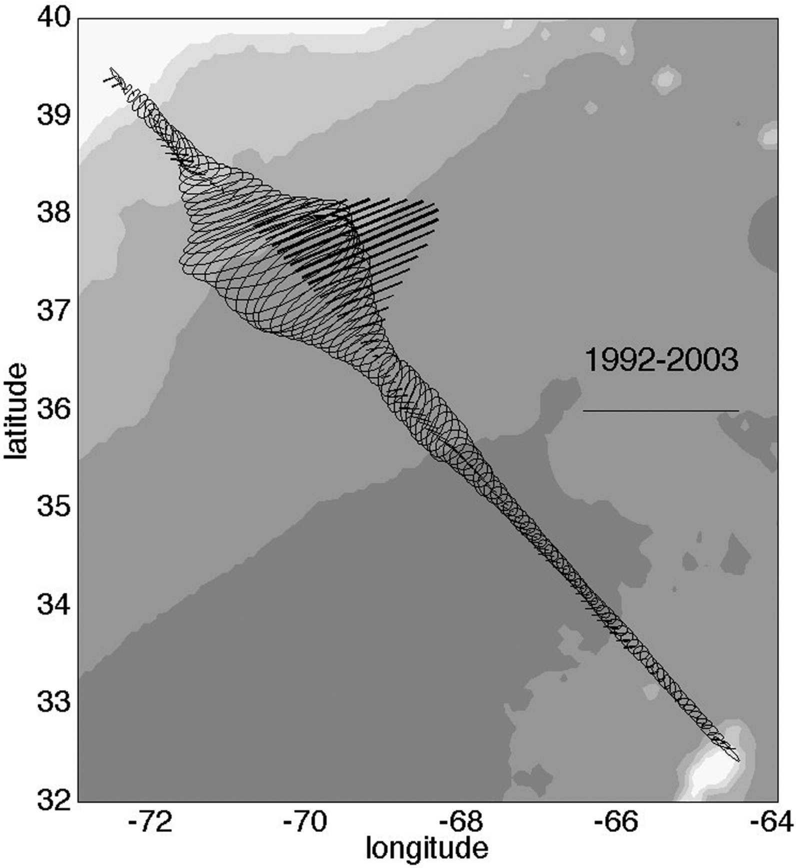

Figure 2. The mean velocity and variance fields from all Oleander ADCP data at 52 m depth between 1992 and 2003. The bar indicates 0.5 m s^{-1} and 0.5 m^2 s^{-2} mean velocity and variance respectively. The depth contours range from less than 1000 to greater than 5000 m.

along the same line. The 'beady' character of the peak speed in the current reflects the effect of meandering on the quarterly averaging of the current. Somewhat more than 200 km south of the Gulf Stream velocity maximum, one finds a band with speeds almost always to the west, and roughly 350 km south of the velocity maximum the currents often are to the east again. The westward flow is nearly always present whereas the eastward flow is more interrupted. The latter extrema often reflect the westward passage of cold core rings across the Oleander line. Examination of individual ring passages shows that swirl velocities often exceed 1 m s^{-1}. Note how the locus of ring crossings follows that of the Gulf Stream: As the stream shifts to the north, so do the rings, they tend to stay close to the current (cf. Richardson, 1983). South of the ring corridor the waters appear to flow more consistently to the west. This can be seen in Figure 4, which shows the mean flow from Figure 3 averaged in two different ways, as a geographical average, and averaged relative

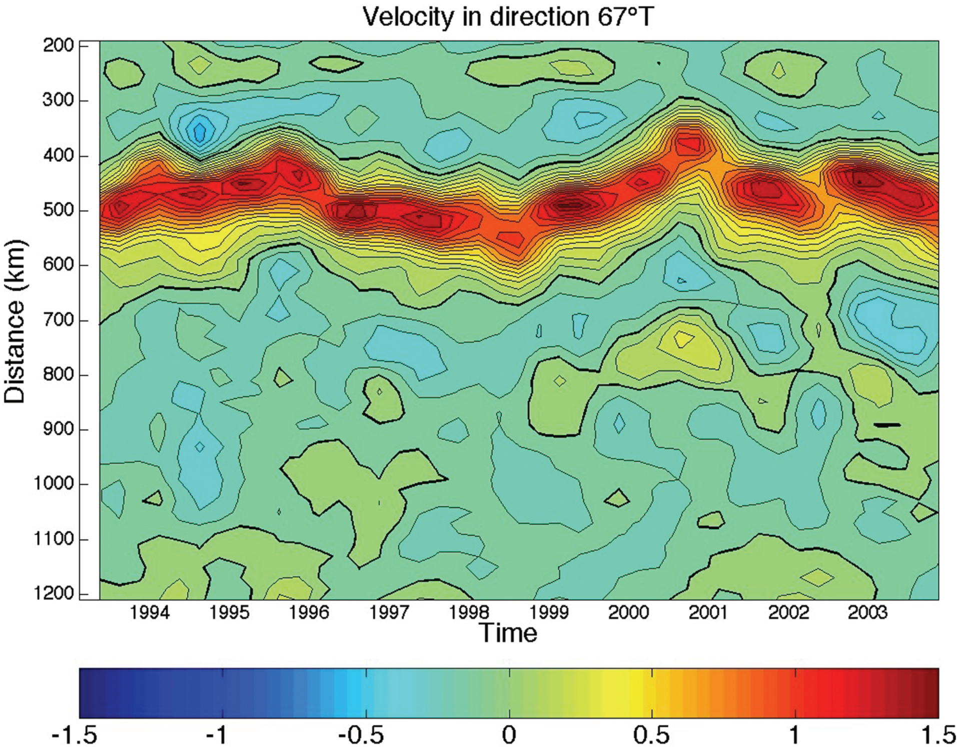

Figure 3. A Hovmöller diagram of velocity between the continental shelfbreak and Bermuda in the direction 67°T corresponding to the mean direction of the Gulf Stream. The heavy lines denote the transition between positive and negative velocities.

to the location of the Gulf Stream velocity maximum. The abscissa plots distance from NJ in both cases, but for the latter the averages from each quarter have been summed relative to the mean position of the Gulf Stream at the time. Both curves show bands of west- and eastward flow, but these become more pronounced in Gulf Stream-locked case (heavy curve): Immediately south of the Gulf Stream the mean flow averages -0.17 m s^{-1} (i.e. to the west) and another 75 km to the south the mean flow is all but zero. Yet another 75 km farther south it reaches—0.09 m s^{-1}, beyond which it gradually diminishes in magnitude toward zero near Bermuda.

Figure 3 also reveals the passage of warm core rings north of the Gulf Stream (at about 250 km from NJ). Here, however, the presence of the continental slope prevents the rings from shifting as freely in response to the meandering of the stream. Thus the west-east pattern of flow from south to north across the Slope Sea shows up quite clearly when averaged geographically in Figure 4 (thin line) and averages close to zero in the 230 to 250 km range. Nonetheless, the westward flow from the southern side of the rings shows up more strongly when plotted relative to the stream (heavy line). The intense westward

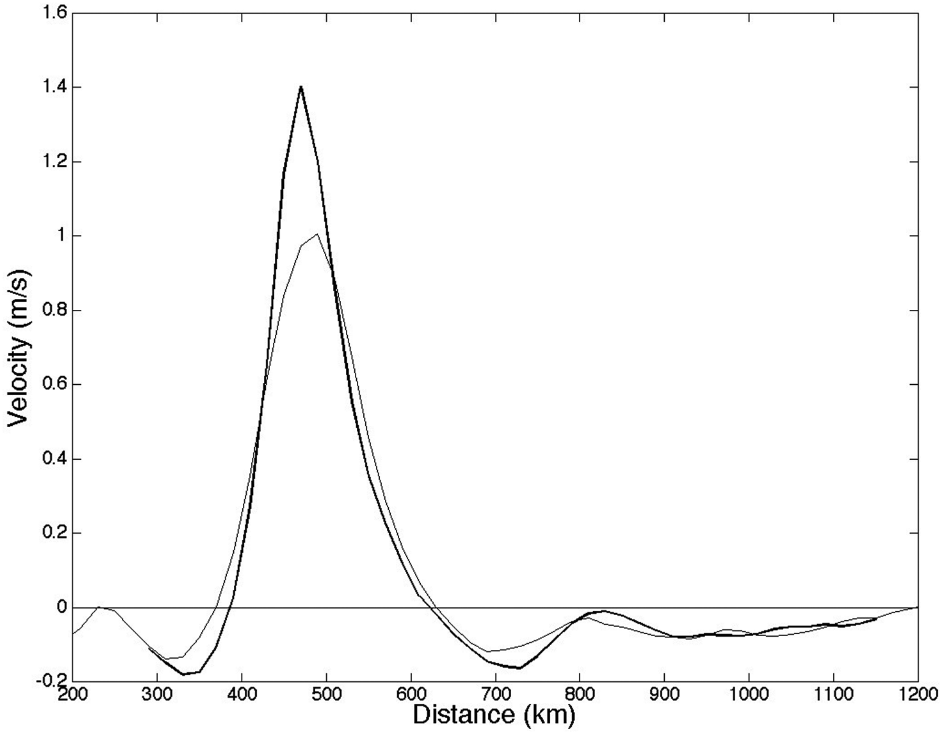

Figure 4. The mean velocity in direction 67°T from Figure 3 plotted as a function of distance (thin line) and as a function of distance after sliding each quarterly segment in Figure 3 so that the maximum velocity of the Gulf Stream sits in the 11-year mean 20 km bin (heavy line).

flow seen in the 3rd quarter of 1994 north of the stream in Figure 3 results from the repeated sampling of a near-stationary warm core ring.

5. Transports

a. The 11-year mean

A major objective of the Oleander program has been to quantify transports in the North Atlantic with particular interest in the Gulf Stream. As before we subdivide the Oleander line into three subregions: the Slope Sea from the continental shelf edge to the Gulf Stream, the Gulf Stream itself, and the Sargasso Sea from the Gulf Stream to Bermuda. Since the Oleander ADCP does not reach to depths that allow for an integration of total transport, the analyses focus on a single shallow layer, defined as one-meter thick at 52 m depth, the mid-depth of the 5th bin of the velocity profiles. While this transport definition reflects an unavoidable instrumental limitation, the Oleander line provides for high-resolution horizontal coverage and frequent sampling of currents. Figure 5 shows the annual averages of the regional integrals in $m^2\ s^{-1}$. The line 'total transport' gives the grand sum of all three.

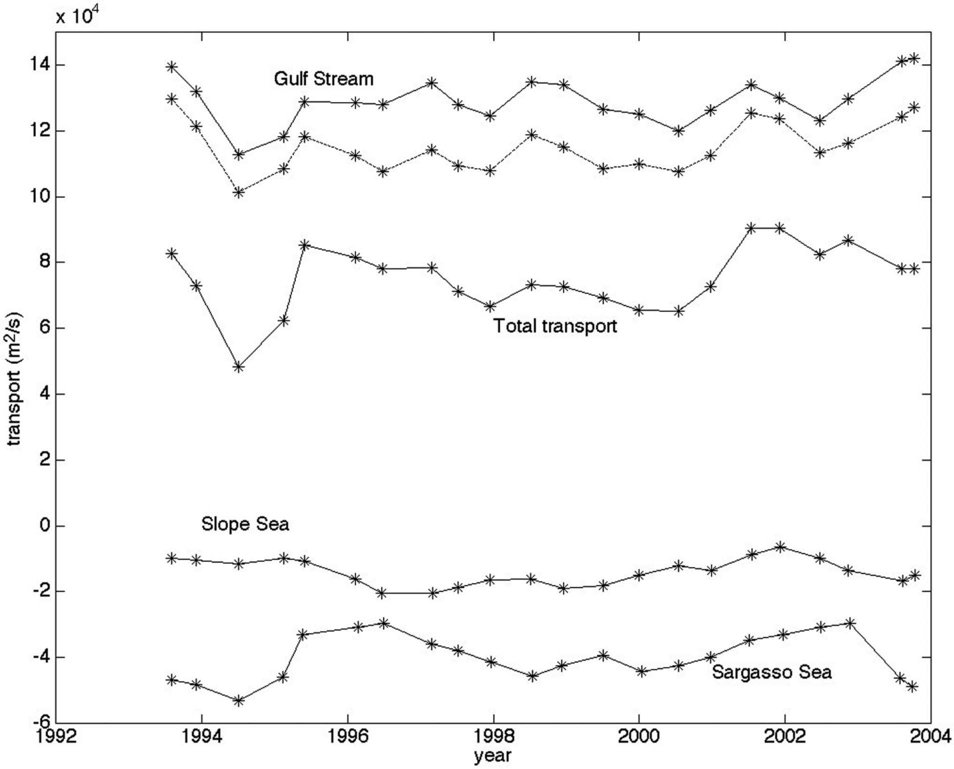

Figure 5. Layer transport in $m^2\ s^{-1}$ at 52 m depth in the Slope Sea, Gulf Stream, Sargasso Sea, and the sum of the three (Total transport). These are computed as 1-year averages every 6 months. The dashed line shows the sum of the Slope and Gulf Stream transports.

Dividing the mean of this, $7.6 \times 10^4\ m^2\ s^{-1}$, by the length of the three sections, 1000 km, gives us a mean velocity normal to the line (west to east) of $0.076\ m\ s^{-1}$.

The mean transports plotted every six months in Figure 5 use a twelve-month window (a factor two overlap). Except for the last year, the six-month steps are almost evenly spaced. The larger step in 2003 results from data loss due to a several month-long subtle compass repeater failure. The changes over time take place rather gradually suggesting some robustness to the estimates, but the noise goes up sharply if the time interval is decreased to nonoverlapping six-month or shorter steps. This reflects the loss of degrees of freedom in averaging over the meandering activity of the Gulf Stream.

The 11-year mean transport in a one-meter thick layer by the Gulf Stream (in stream-coordinates) is $12.8 \times 10^4\ m^2\ s^{-1}$. It is appropriate at this point to compare this transport with corresponding estimates from other studies, in particular the Halkin and Rossby (1985) Pegasus data set from 73W. To do this, the mean Gulf Stream velocity structure in stream coordinates, as reported by Halkin *et al.* (1985), was used to determine

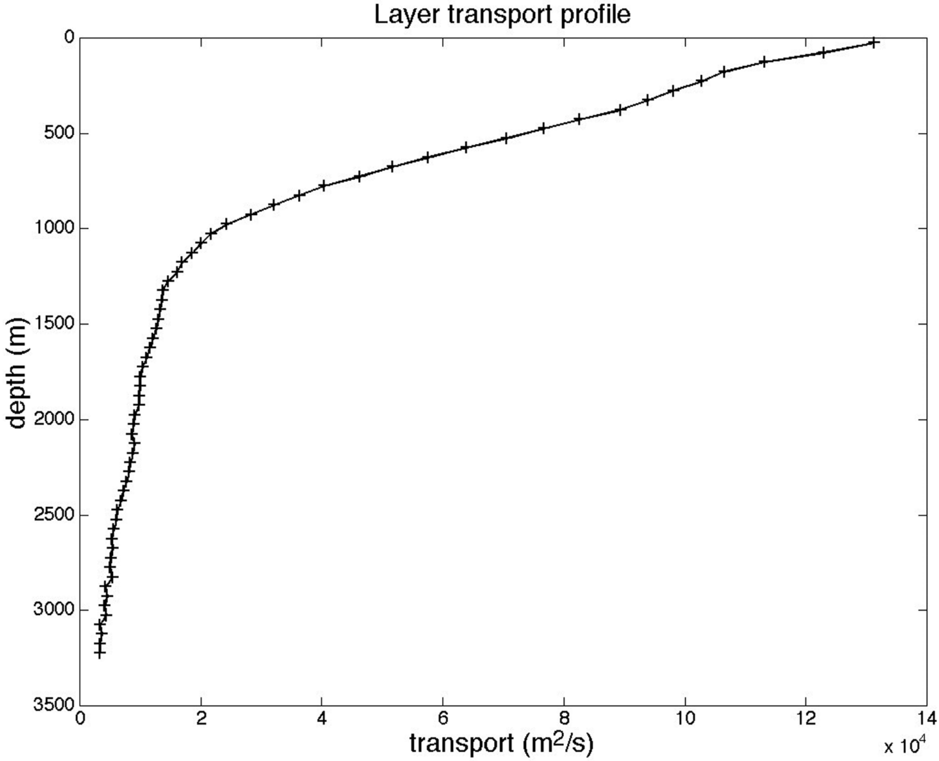

Figure 6. Profile of layer transport in $m^2 s^{-1}$ estimated from the mean Gulf Stream in stream coordinates using the 3-year Pegasus data set (Halkin and Rossby, 1985). The transports are estimated for every 50 m starting at 25 m depth. The total transport from the surface to 2000 m equals 87 Sv and 94 Sv to 3250 m. The baroclinic transport using 2000 m as a reference layer = 69 Sv.

layer transport as a function of depth. This was done by integrating across the current from the surface to 3250 m (total water depth ~ 3500 m) in 50 m steps. The resulting layer transport profile, Figure 6, shows clearly the baroclinic shear of the upper ocean and its transition to the deeper much less sheared deep velocity field below 1000 m depth. Given that the profile exhibits considerable shear even near the surface, it seems quite remarkable that at 50 m depth the layer transport at $12.5 \times 10^4\ m^2\ s^{-1}$ is within 3% of that of the 11-year Oleander estimate above. The mean velocity field from the Pegasus data set is based on about 16 separate or independent sections across the current.

A geostrophic estimate of mean transport near the surface can be obtained from the Sato and Rossby (1995) mean dynamic height difference across the stream relative to 2000 m: $d\Delta D = \Delta D_{\text{Sarg}} - \Delta D_{\text{Slope}} = 2.26 - 1.21 = 1.05$ dyn m at the surface (from their Table 1). By integrating in the cross-stream direction the geostrophic balance, $-fv = -g\partial\eta/\partial x$, where v is downstream velocity, x cross-stream distance, f the Coriolis parameter, η surface elevation, and g acceleration due to gravity, we obtain the layer transport $V = \int v dx = (g/f)\Delta\eta$, where $\Delta\eta$ is sea level difference from one side to the other defined by where the downstream velocity drops to zero. Interpreting the dynamic height difference as sea level change, $d\Delta D = (g/10)\Delta\eta$, we obtain $V = (10/f)\ d\Delta D = 1.17 \times 10^5\ m^2\ s^{-1}$. Using the transport profile in Figure 6 to project this surface value to 50 m depth, we obtain $(1.17 - 0.10) \times 10^5 = 1.07 \times 10^5\ m^2\ s^{-1}$. Since this includes only the baroclinic portion of transport, the transport at 2000 m depth (from Fig. 6) is added to obtain a total layer transport of $(1.07 + 0.09) \times 10^5 = 1.16 \times 10^5\ m^2\ s^{-1}$. Whereas the two direct measurements, ADCP and Pegasus profiles, agree rather closely, this geostrophic estimate, after adjusting for shear and adding the barotropic component, yields a 10% lower estimate. Almost certainly this underestimate reflects, at least in part, the fact that most hydrographic sections across the GS provide only limited horizontal resolution of the current including where the sign of the current changes due to *the return flows* to either side (as in Fig. 4). As a result, unless two stations bracketing the stream are taken right at the locations of dynamic height maximum and minimum south and north of the current, respectively, the dynamic height difference and hence the transport must have a negative bias.

The Slope Sea and Sargasso Sea both have mean flows to the west, as expected. Dividing the mean Slope Sea transport, $1.37 \times 10^4\ m^2\ s^{-1}$ by its 200 km mean width yields a mean velocity of $0.069\ m\ s^{-1}$. The corresponding mean velocity in the Sargasso Sea is $3.82 \times 10^4\ m^2\ s^{-1}$ divided by 600 km or $0.064\ m\ s^{-1}$, or almost the same as in the Slope Sea.

b. Interannual variations

With respect to interannual variations plotted in Figure 5, the Slope Sea clearly shows the least transport change in absolute terms although in relative terms Slope Sea transport can vary by a factor 2. In the 1996–1998 time frame, coincident with a southerly position of the Gulf Stream (Fig. 3), the westward transport was nearly double that when the stream was displaced to the north. Curiously, the fluctuations in Slope Sea transport, although substantial, appear to vary rather gradually, as if heavily low-pass filtered, perhaps reflecting its constrained geometry between the Gulf Stream and the shelfbreak. What causes the Slope Sea circulation to vary over time is of considerable interest and will be discussed below.

The Gulf Stream transport (top curve) is of course the largest contributor to transport between the continental shelf and Bermuda. The standard deviation of the time series is only 6% but a very large dip in transport occurred in 1993–1995. A variety of time scales can be seen with what would appear to be a general increase in transport during the last four years, but viewed overall there is no suggestion of a longer-term trend in the record. Knowing that the Slope Sea transport must turn back to the east near Cape Hatteras, the dashed line in Figure 5 shows the sum of the Slope Sea and Gulf Stream transports, a number that might provide a more accurate picture of net eastward flow. The effect is

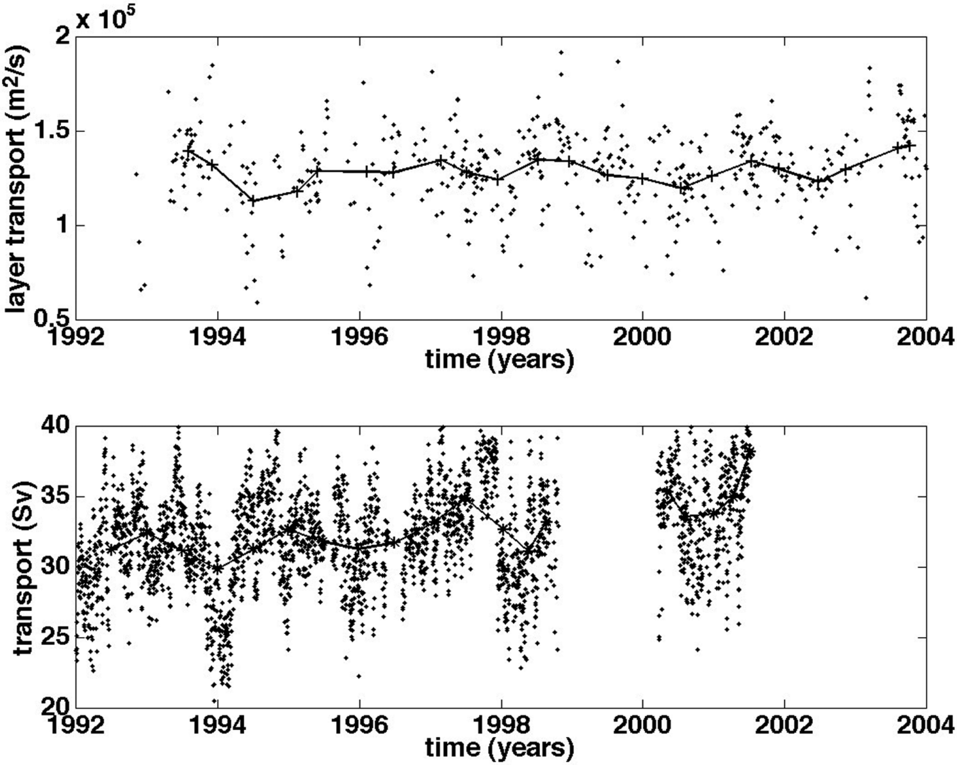

Figure 7. Gulf Stream layer transport as in Figure 5 (top panel). Transport in the Florida straits for the same period as determined by electromagnetic methods (bottom panel). The annual averages were computed the same way as for layer transports. The Florida data were obtained from http://www.pmel.noaa.gov/wbcurrents/. The reader is encouraged to consult this site for the specifics about these measurements.

rather minor, however, leaving one with the impression of a range of time scales, none particularly dominant within this 11-year record.

The 600-km wide Sargasso Sea section exhibits a nearly factor three range in transport principally due to strong westward flow in 1994 and subsequent decrease in 1995. This event occurs early in the Oleander program raising the concern that it might be an instrumental artifact (specifically a systematic compass error); after all it shows up in the Gulf Stream transport too. If this were the case, it would seem plausible to expect a similar bias in the Slope Sea transport; but quite to the contrary, there is absolutely no hint of this signal there. Interestingly, two other time series provide suggestive evidence that this event in 1994–1995 may be real. First, the long running and sustained program of cable measurements of transport in the Florida Straits (see Baringer and Larsen, 2001 for a nice overview). The measurements show a distinct event at the start of 1994. This can be seen in the lower panel of Figure 7, which for direct comparison also includes the ADCP-based

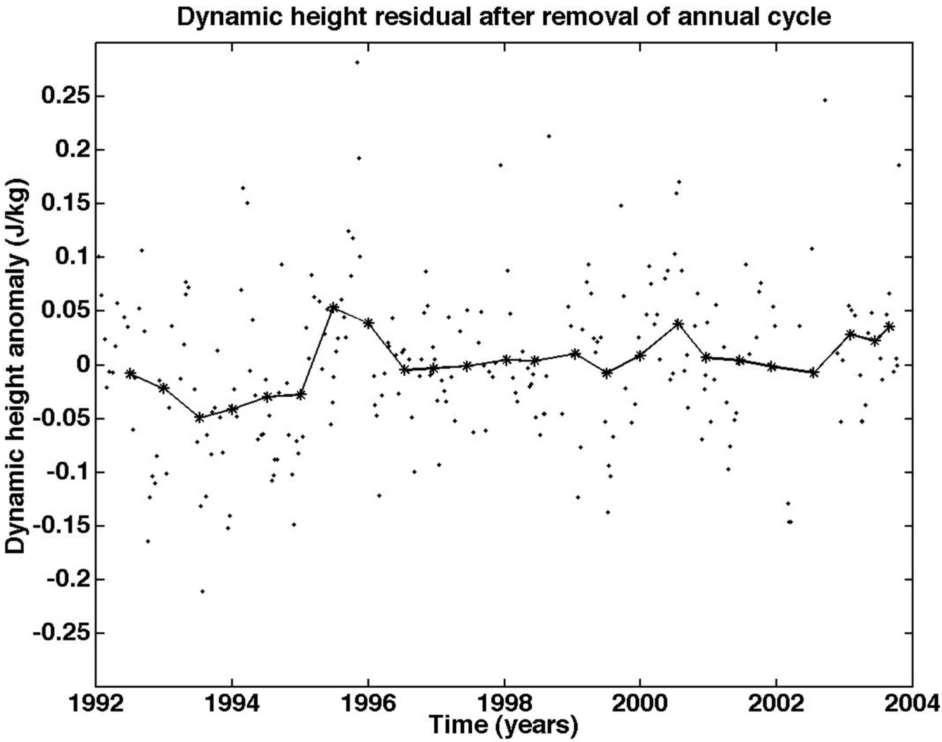

Figure 8. The 0–1500 dbar dynamic height residual at Station 'S' after removing the annual cycle and mean for the 1992–2003 record. The annual averages were computed the same way as for layer transports.

Gulf Stream record in the top panel. Actually, the daily cable data (dots) show the event more sharply than the running yearly averages (estimated as above). There also appears to be a slight lag in the event from Florida to the Oleander line. Another event that appears to show up in both records occurred near the start of 1998.

The long-running hydrographic time series at Station 'S' at Bermuda provides the other line of evidence. Figure 8 shows the 0–1500 dbar dynamic-height anomaly after removing the mean and annual cycle from the 1992–2003 data set. It shows a drop in dynamic height anomaly in 1993 followed by a striking increase in early 1995. An increase in dynamic height anomaly reduces the difference between Bermuda and the southern edge of the Gulf Stream implying a decreased westward geostrophic transport. A 10 dyn cm increase would lead to a mean transport drop of $g \times 0.1/f(\sim 35\text{N}) = 1.2 \times 10^4\ \text{m}^2\ \text{s}^{-1}$ or roughly one half the jump seen in Figure 5.

The fact that the Florida Straits cable transport and Bermuda hydrographic record both showed an event at about the right time and with the right signs suggests that the large westward transport in the Sargasso Sea in 1994 is real and not due to instrumental error or limitations. Curiously, the rest of the time series in the Sargasso Sea, the broad increase in transport in the late 1990s followed by a gradual decrease appears to mimic that of the Slope Sea, both in trend and amplitude, but with a slight delay. But this slow pattern spans the better part of a decade rendering any conclusion about its significance highly uncertain.

6. Inferring total transport in the Gulf Stream from layer transport

As noted earlier, technical factors limit the reach of the Oleander ADCP to only modest depths, 200–300 m in the Slope Sea and roughly 150 m in the Sargasso Sea. This is a definite handicap; one would like to profile to much greater depths as a matter of routine, and this will change with the new ADCP that has recently been installed on the Oleander. But it had already been established in the early 1980s (Halkin and Rossby, 1985) that the Gulf Stream exhibits considerable stiffness, both horizontally and vertically. Rossby (1987) showed that nearly 80% of the Eulerian eddy kinetic energy in the center of the Gulf Stream's meandering envelope can be accounted for in terms of a strictly stiff current with meandering as its only freedom of movement. This stiffness of the current provides a possible framework for mapping the observed layer transport into total transport, the quantity of real interest. A difficulty here is that, and as Figure 6 shows, the Gulf Stream consists of an upper-ocean baroclinic and a deep-ocean barotropic component, both of which will contribute to upper-ocean layer transport. Unfortunately, the Oleander program does not include a monitor of deep-ocean transport (or pressure difference). Lacking further information it will be assumed that all variability is baroclinic with no variability in the deep waters. On meandering time scales, bottom pressure can vary $O(10)$ dyn cm, but at annual periods variations in cross-stream pressure difference very likely lie in the $O(1)$ dyn cm or less range. But the variability of the deep pressure field is one that needs to be studied and understood much better.

Given a baroclinic Gulf Stream, a simple relationship between layer transport and total transport can be developed. It has been shown that the vertical structure of the current at each and every point across the current closely resembles the locally estimated gravest flat-bottom baroclinic mode (Rossby, 1987). For the purposes of this study, this suggests a further simplification, namely to use the classic reduced-gravity model to obtain both layer transport and total baroclinic transport. Using the standard notation for such a model where the interface shoals (but does not surface) to the north, the y-momentum equation is $fu = -g'\partial h/\partial y$. Integrating across the current, the transport for a layer near the surface becomes $T_{\mathrm{sfc}} = -(g'/f)\int \partial h/\partial y dy = -(g'/f)(h_+ - h_-)$ where the h_+ and h_- represent upper-layer thickness on the northern and southern sides, respectively. The corresponding upper-layer transport can be obtained similarly: $T_{\mathrm{tot}} = -(g'/f)\int h\partial h/\partial y dy = -1/2(g'/f)(h_+^2 - h_-^2)$. If we assume that fractional variations in $(h_+ + h_-)$ are less than in $(h_+ - h_-)$ we obtain the simple relationship $T_{\mathrm{tot}} = \langle h \rangle T_{\mathrm{sfc}}$ where $\langle h \rangle$ is the average of the two depths. The principal assumptions made include no variation in the deep field on the annual and longer time scales of interest here, and that variations in upper-layer transport are small and result from pycnocline depth variations, not density variations. The pycnocline itself has been reduced to a sharp interface between two layers. No assumption need be made about the shape or width of the current. Using typical values for g' and f (0.02 m s^{-2} and 0.9×10^{-4} s^{-1}), $h_+ = 300$ and $h_- = 900$ m, $T_{sfc} = 1.31 \times 10^5$ m^2 s^{-1}, and $T_{tot} = 78$ Sv. These fluxes agree very well with observed values from both the Oleander and the earlier Pegasus programs.

The slope of the curve in Figure 6 indicates the rate of increase in layer transport due to the shoaling pycnocline. Holding the offshore or southern layer thickness h_- at 900 m and letting the shallow side h_+ shoal, it follows that $dT_{lyr}/dz = -(g'/f) \int \partial^2(h_+ - h_-)/\partial y \partial z dy = -(g'/f)\partial(h_+ - h_-)/\partial z = -(g'/f)(\partial h_+/\partial z - 0) = (g'/f) = 220$ m^2 s^{-1} per m. This is quite a bit higher than that observed in Figure 6 and may be due to the over-simplified representation of a continuously stratified pycnocline as a sharp interface. In summary, under the assumptions made above, variations in layer transport by the Oleander can be mapped into variations in baroclinic transport when multiplied by the mean ~600 m depth of the pycnocline. But more important than the precise multiplier, this result says that the *fractional* range of variability for the total baroclinic transport should be about the same as for the observed layer transport. This is important for it says that the 6% standard deviation in layer transport, when averaged annually, also applies to the total baroclinic transport.

7. Discussion

The uninterrupted monitoring of upper-ocean currents (now in its 13th year) by the Oleander is opening up new windows of study regarding low-frequency variability of the Gulf Stream and surrounding waters in the northwest Atlantic. Whereas in the past estimates of fluxes were indirect and infrequent, the strength of this program lies in the repeat and accurate measurement of currents over a wide range of horizontal scales, a dimension that until recently has been rather inaccessible to study. This paper has used these data to examine the variability of upper-ocean fluxes on interannual time scales.

The most important improvement that results from the frequent sampling of transport is of course the reduction of high-frequency variability through the averaging process. The top panel of Figure 7 illustrates this point. It shows all measurements of Gulf Stream layer transport by the Oleander over the eleven years. The superimposed black line shows annual estimates computed semiannually as plotted in Figure 5. The standard error associated with the annual estimates is about 3% compared to the nearly 20% scatter in the data. Thus the $O(6)\%$ interannual variations in Gulf Stream should be seen as significant. But the Oleander line also samples the Slope and Sargasso seas and a very striking feature about these three subregions is their different temporal behavior. Change in Slope Sea transport exhibits a rather gradual character whereas in the Sargasso Sea fluxes can change quite rapidly. Further, there is a 'ripple' in the Gulf Stream flux with a repeat pattern around two years that does not appear to be present in either the Slope or Sargasso seas. Whether this 'ripple' reflects variability due to the meandering of the current, or has a larger-scale wind-driven origin remains to be determined. In any event, each of the three time series of transports have distinctive characteristics suggesting partially different dynamics at work in the different regions providing further evidence that the observed patterns of variability are real and not the result of instrument or measurement uncertainties.

a. The Gulf Stream

Considering first the Gulf Stream subset, it has a mean layer transport at 52 m depth of $(12.8 \pm .72) \times 10^4\ m^2\ s^{-1}$, a rate that projects to a total transport of ~78 Sv using the 600 m scale factor developed above. The observed 6% standard deviation maps into a transport variability of 4.6 Sv while the corresponding peak-to-peak range of $(11.3$ to $14.2) \times 10^4\ m^2\ s^{-1}$ implies a 23% or rather substantial 17 Sv range with a minimum in mid-1994 and maxima in 1993 and 2003. The Gulf Stream transport seems to exhibit the widest and liveliest range of time scales, as can be seen in Figure 5. These substantial variations in transport on rather fast time scales, of no more than a couple of years, are so far as we know, new results since none of the earlier Gulf Stream programs had the sampling density to allow a direct comparison.

Earlier studies of Gulf Stream transport group into two general classes, one that considers the Gulf Stream itself and one that views the Gulf Stream more broadly within the North Atlantic. The most relevant one of the first group is that of Sato and Rossby (1995), a study that follows a tradition going back to Worthington (1976) and Iselin (1940). They used 130 hydrographic sections from the 1930s through the 1980s to search for long-term variations in baroclinic transport. Within the limitations set by the 130 sections spread over five decades, nothing could be said about interannual variations, but at the pentadal scale they noted some indication of an increase at the $O(4)$ Sv level between the first and second half of the 1970s. There was no difference between the late 70s and late 30s, the only other reasonably well-sampled pentad (20 sections). If one tries the same 'game' of splitting the Oleander record into two pentads (5.5 years each), it turns out that these two periods agree to within 1%. However, it should also be noted that when Sturges and Hong (2001) extended their previously developed wind-driven model of sea-level variations (Sturges and Hong, 1995) to estimate sea level difference across the Gulf Stream off Norfolk, Virginia, they found that they could indeed account for the pentadal variations in transport reported by Sato and Rossby (1995).

The other class of Gulf Stream transport studies focuses on a larger scale. The first of these, discussed earlier, is that by Greatbatch *et al.* (1991), but they consider only two pentads fifteen years apart, not what happens within the pentads. More germane to the Oleander program is the study by Curry and McCartney (2001) in which they use the hydrographic time series at Bermuda and the center of the Labrador Sea (ocean weather station 'Bravo') to construct a potential energy anomaly record for the two sites, the difference between which provides a measure of baroclinic transport (Fofonoff, 1962). They call this quantity a transport index in analogy with the North Atlantic Oscillation (NAO) Index. Consistent with the Greatbatch *et al.* (1991) study, they find a 10 Sv decrease in baroclinic transport between the late 1950s and early 1970s. They also observe an increase of 15 Sv between 1970 and the mid-1990s, a range in transport comparable to the extrema observed here. Unfortunately, there is a fundamental difference between their study and this one that makes direct comparison not only difficult but meaningless. The reason is that their two-point potential energy anomaly difference must, by definition, include currents on all scales anywhere between the central Labrador Sea and Bermuda. This is particularly true of the ocean because the dynamically active scale is set by the radius of deformation, which is only $O(10\text{–}35)$ km compared to the $O(10^3)$ km for the atmosphere, a scale more consistent with that upon which the NAO index is based.

b. The Slope Sea

The mean upper-layer transport to the west varies between $(0.66 \text{ vs. } 1.97) \times 10^4 \text{ m}^2 \text{ s}^{-1}$ with a mean of $1.37 \times 10^4 \text{ m}^2 \text{ s}^{-1}$. Because the baroclinic shear is so much weaker in the Slope Sea, there is very little change in layer transport from 52 to 252 m (but as the mean vectors in Figure 2 show, there is quite a bit of horizontal shear *across* the Slope Sea). Thus, the corresponding transport assuming no shear in the top 252 m would be about 3.5 Sv to the west. As noted earlier, layer transport in the Slope Sea varies only gradually such that the record to date appears to be dominated by little more than one complete cycle. The range, while numerically smaller than for other regions, varies nonetheless by a factor 2.

Between 1996 and 1999 the transport was double that in 1994 and 2000. Further, during this period the Gulf Stream also assumed a significantly more southerly position (see Fig. 3) and the surface salinity and upper-layer temperatures were anomalously fresher and colder than usual (Fig. 9). This apparent correlation had been noted earlier by Rossby and Benway (2000), who suggested that the lateral shifting of the Gulf Stream might be part of a dynamical response to the increased (re-)circulation of waters in the Slope Sea. They also noted that as the westward flux increases, the surface salinity of the Slope Sea waters decreases significantly, the reason being that with the greater Slope Sea transport it follows that there would be a greater dilution of the more saline waters continually leaking from the Gulf Stream (see also Schollaert *et al.,* 2004). They suggested a correlation with the NAO index such that at times of high NAO (cold and windy in the Labrador Sea) less fresh water will flow south past the Grand Banks due to a reduction in cross-shelf pressure gradient from the freezing and production of dense salty waters. In a recent study Hameed and Piontkovski (2004), using Gulf Stream northwall position data from Taylor and Stevens (1998), note that rather than the NAO index, the lateral shifting of the path of the Gulf Stream on interannual time scales correlates much better with the intensity and position of the North Atlantic low. Their conclusion suggests a thermohaline or buoyancy-related forcing of the path of the Gulf Stream: when the Icelandic low is weak and positioned farther to the east, the Labrador Sea will experience less convective cooling and densification of surface waters than otherwise. These relatively fresh upper ocean waters are then exported south and west along the continental shelfbreak into the Slope Sea where they both 'push' on the Gulf Stream and also must turn back east since they are unable to cross underneath the Gulf Stream to the south (Bower and Hunt, 2000; Hogg and Stommel,

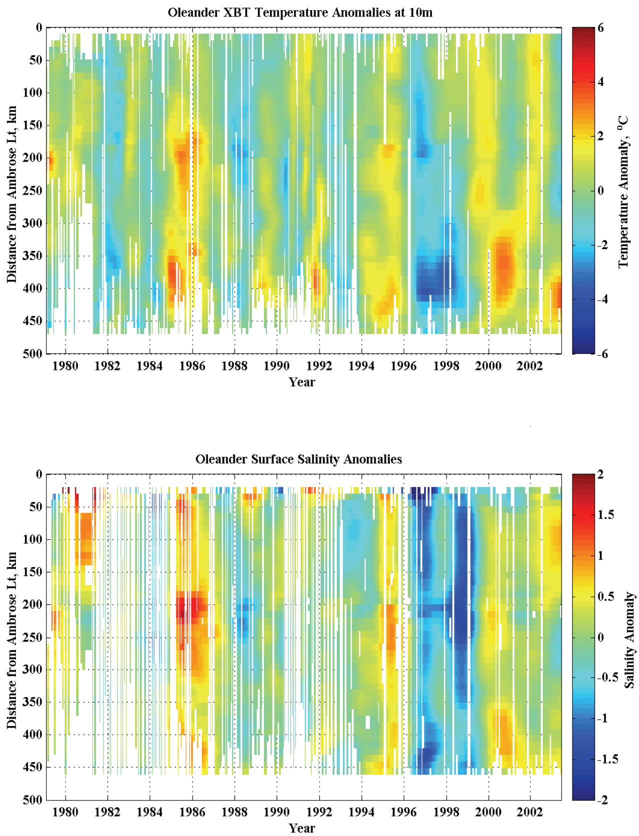

Figure 9. These two Hovmöller diagrams show near-surface temperature (top) and salinity (bottom) anomaly across the continental shelf and Slope Sea from the NOAA/NMFS XBT program on the Oleander. The anomaly is the residual after removing the annual cycle from the data (Rossby and Benway, 2000). The data have been low-passed filtered with a 6-month half-power point.

1985). In cold winters, more of these waters will be cooled and transformed into dense North Atlantic Intermediate waters that are exported primarily into the eastern Atlantic at 1500–2000 m depths (Bersch *et al.,* 1999), but also south and west toward the Gulf Stream (Joyce *et al.,* 2000).

But there is another, very different view of what controls the latitude of the Gulf Stream, or *more accurately* the latitude at which it separates from the continent. It is beyond the scope of this paper to discuss in detail the Parsons-Veronis hypothesis on the Gulf Stream separation, but the theory holds that separation must take place at the latitude at which the thermocline depth difference across the Gulf Stream reaches a maximum (surfaces on the inshore side). This idea was examined in detail by Gangopadhyay *et al.* (1992). They found a best correlation between separation latitude and the integrated southward Ekman layer transport across the Atlantic when the latter was integrated for about three years, a time scale somewhat less than the estimated transit time for long Rossby waves.

It might seem that these two quite different ideas must be contradictory, but this need not be the case. The separation latitude itself may well be determined by the dynamics of the western boundary current, i.e. the Parsons-Veronis hypothesis and as further examined in the Gangopadhyay *et al.* study. But there is nothing in the theory that precludes other processes such as thermohaline forcing from coming into play after separation has taken place. This may be an area ripe for further study.

c. The Sargasso Sea

The Sargasso Sea layer transport record to date exhibits a curious mixture of time scales, both rapid change and gradual drift: From a maximum of 5.4×10^4 m^2 s^{-1} in 1994 it drops to 3.0×10^4 m^2 s^{-1} by late 1995, a change that would imply a transport decrease of $O(19)$ Sv for an 800 m upper layer baroclinic flow. Curiously, the fast event in 1994–1995 also appears in the Gulf Stream (with comparable magnitude and sign) but not in the Slope Sea. The Sargasso Sea also shows a rapid increase in transport in 2003. These rather quick changes suggest a wind-driven rather than a buoyancy-driven mechanism, particularly given the Sargasso Sea's location within the wind-driven subtropical gyre. In a very telling study, Sturges and Hong (1995) were able to reproduce the dynamic-height variations at Bermuda using a linear, stratified model of the North Atlantic Ocean forced by NCEP-reanalyzed winds. In particular, they could reproduce the amplitude of the 1970 dynamic-height minimum at Bermuda with astonishing accuracy. While currents and dynamic height cannot be compared directly (without other information), the fact that the 1994 transport event shows up in the Bermuda hydrographic record suggests a wind-driven origin to this event.

The long time-scale drift in the Sargasso Sea transport would appear to follow the Slope Sea trend with a year or two's lag, but whether this has physical significance remains to be determined. For example, they may be related in a general atmospheric sense, but driven by quite different physics, the Slope Sea primarily through buoyancy and the Sargasso Sea through mechanical forcing.

One more question: why did the 1994 fast event in both the Gulf Stream and the Sargasso Sea not show up in the Slope Sea? In both cases the drop is of the same sign. Why did it not show up as an increase in flow to the west in the Slope Sea? Had it done so, we might have been more concerned that it was instrumental, notwithstanding that it shows up in two other geophysical records. The only suggestion that comes to mind is that the flow in the Slope Sea is tightly locked to the topography along which it flows and driven by the 'strength of its source' in the Labrador Sea. It appears to be insensitive to what happens locally in the western North Atlantic.

d. Heat transport

A question of considerable importance concerns variations in heat transport associated with the observed layer and volume flux variations. This, unfortunately, is a difficult problem because concurrent observations of temperature do not exist. But one can put certain bounds on the heat or more accurately, the temperature transport, given knowledge of variations in volume flux. The distinction between heat and temperature flux is that the latter is not subject to the zero mass flux constraint. But the sum of all temperature fluxes across a surface of no-net mass flux will yield the (net) heat transport for that section. Besides the practical necessity of working with temperature fluxes (different types of data in different regions), these partial integrals can be useful for examining sources and magnitudes of variability.

What might transport variations in the Gulf Stream imply about changes in heat flux? There are two steps one can take to simplify this question. First, except for the surface waters in direct contact with the atmosphere, the temperature—salinity—density relationship remains essentially constant. Any change in temperature transport will depend upon the speed of a parcel and whether the volume of a particular density class changes. Second, it appears that the thermocline rises and falls with little change in shape such that stratification and hence the layer thickness of the various density classes remains the same. Given this, it follows that temperature transport by the waters in the main thermocline will vary linearly with volume transport. Sato and Rossby (2000) estimate the total temperature transport for a 73 Sv baroclinic Gulf Stream at 4.84 TW (teraWatt) with a (net) heat flux across 36N latitude of 1.2 TW. If one considers the 6% transport variations in the Gulf Stream as typical, they would imply temperature flux variations of about $0.06 \times 4.84 = 0.3$ TW. But a good fraction of this is undoubtedly recycled such that the net heat flux variations will be much smaller. How much is unclear, but it would be about 0.075 TW if the net perturbations take place in the same proportion as the net-to-total heat flux. Interestingly, one obtains a similar heat flux variation if one lets surface temperature vary while holding the upper layer volume flux constant. For illustration, consider a temperature change of 1°C for the top 100 m. This corresponds to a heat flux perturbation of 0.05 TW. These may seem like rather small numbers compared to the ± 0.3 PW uncertainty and 0.6 PW range of the poleward heat flux estimates by Sato and Rossby (2000). On the other hand, if one uses the 23% range (from Fig. 5) and 4°C range in SST anomaly (top panel of Fig. 9 below), the corresponding temperature fluxes could imply significant variations in poleward heat flux.

A different type of heat flux, and one with perhaps greater atmospheric impact, concerns the consequences of lateral (or latitudinal) displacements of the path of the Gulf Stream. When the Gulf Stream shifts to the south, presumably in response to a greater westward transport in the Slope Sea as discussed above, there is a concomitant decrease in temperature and salinity of the surface and near-surface waters of the Slope Sea. So, in addition to a drop in SST due to the shift of the stream away from its former position, SST drops throughout the entire region north of the stream, i.e. the Slope *and* the shelf waters. And conversely, when the stream shifts to the north—presumably due to a reduction in cold water supply in the Slope Sea—the SST and surface salinities (SSS) increase conspicuously. The range in SST and SSS can exceed several degrees and PSU and the area over which these changes occur is of $O(10^3)$ km east-west and several hundred km north south. Figure 9 shows warm periods around 1985–86, 1994–1995 and rather more broadly since 1999. A striking cold period occurred in 1997–1998 and earlier between 1982–1984. Since Ambrose Light sits at the entrance to New York Harbor and the Gulf Stream starts at or a bit beyond 450 km, these variations in temperature and salinity occur across the entire region, both the Slope Sea and the shelf waters. Between the warm and cold periods the average temperature anomalies may differ by up to 4°C, an enormous difference over a huge area. The fact that these variations occur in tandem with lateral shifts in the Gulf Stream suggests that they result from time-varying transports along the continental margin and not from atmospheric forcing in the region (Rossby and Benway, 2000; Joyce *et al.,* 2000; Hameed and Piontkovski, 2004). This statement can be reinforced by the fact that salinity changes of the magnitude observed cannot possibly be obtained through evaporation/precipitation since to effect a 2 PSU change to a 50 m deep mixed layer with a mean salinity of 35 PSU would require the addition or evaporation of $O(3)$ m of fresh water! Rossby and Benway (2000) also noted that the temperature/salinity anomalies in the Slope Sea tended to be density neutral (or compensating) implying an internal or dynamical constraint rather than merely an atmospheric signature.

8. Summary

Perhaps the single most gratifying aspect about the Oleander ADCP operation has been the cooperation of the vessel owners and operators. Without this interest and willingness to accept our presence on the vessel none of the studies reported here and in other papers would have been possible. Equally important has been that the equipment has been up to the task. Of course it took some time to get all the bugs worked out, but not once during the twelve years of operation was it necessary to take out the ADCP for repair. This kind of reliability has made the long-term operation of this project far more manageable than otherwise would be the case. The ADCP instrument, although limited in its vertical coverage, provides excellent horizontal resolution and coverage, and through continual resampling one can discern quantifiable variations in structure, position, amplitude and persistence of ocean currents all along the route. Already, this methodology is in use on several vessels in the Atlantic and in the Pacific, and it seems more than likely that the coverage provided by the merchant marine fleet will prove attractive to other research initiatives. Vessel operators have been quite supportive of such initiatives as long as they do not interfere with their normal operations.

This paper represents a first effort to focus on the very longest time scales in the 11-year Oleander data set. It has quantified to an extent not possible before the nature and magnitude of ocean currents in different regions. The three regions included in this study, the Slope Sea, the Gulf Stream and the Sargasso Sea, exhibit quite distinct characteristics of behavior. The Slope Sea, despite its immediate proximity to the Gulf Stream, exhibits the least variability in transport with time. Its transport can vary by a factor two in magnitude, but does so only very gradually, as if subject to a severe low-pass filter. In so far as this data set goes, it is the least variable of the three regions. In contrast, the Gulf Stream exhibits the liveliest range of variability, both in terms of amplitude and temporal activity. While the annual averages have a standard deviation of 6%, the range from the lowest to highest annual estimates exceeds 20%. Further, at least the largest of these fluctuations appear to be correlated with changes in transport in the Florida Straits suggesting a large-scale probably wind-driven cause as Sturges and Hong (2001) have shown. The third region, the Sargasso Sea, exhibits a mixture of both fast Gulf Stream and slow (Slope?) characteristics, with a few fast events superimposed on a more gradual variability, one that appears similar to that of the Slope Sea, but lagged by a couple of years. Whether this is significant or not is probably too early to tell.

The high-resolution repeat sampling, in addition to resolving time scales, also reveals 'fine-structure' to the mean field, a mean field best defined in relation to the Gulf Stream. As the stream shifts north and south, cold-core rings do likewise giving the mean field, when referenced to the stream, a distinct banded pattern. The Gulf Stream itself has a width of ~200 km when defined by the zero-crossing of the mean down-stream velocity field.

The position of the stream appears to depend upon the strength of flow in the Slope Sea such that the path of the Gulf Stream shifts to the south during times of large Slope Sea transport, and conversely to the north when the flow diminishes. But the transport *by* the Gulf Stream can vary widely and rapidly, i.e. on quite different time scales. Thus, variations in Gulf Stream position and transport would appear to be governed by quite different processes: position due to Slope Sea transport, perhaps in response to buoyancy-related processes in the Labrador Sea, and transport principally due to changing winds, most likely over the tropics and subtropics. These should be testable hypotheses.

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The research in this paper would not have been possible without the help and support of many. Mr. Cor Teeuwen was absolutely central to getting the Oleander instrumented in the first place. The ship's officers have been very gracious to keep an eye on the equipment and contact us in case anything seemed amiss. The dedication of Mr. George Schwartze to setting and maintaining a high standard of operation for the entire period of operation is truly extraordinary. Dr. Erik Gottlieb played a central role in developing the data processing routines and analysis in the early years of the project. Ms. Sandy Fontana provided crucial knowledge and assistance with the data processing. The ADCP program has benefited enormously from the support and cooperation by Mr. Robert Benway of the National Marine Fisheries Service. Mr. Benway has been in charge of the XBT, sea-surface salinity (the data in Fig. 9) and continuous plankton recorder programs on the Oleander and predecessor vessels for over 25 years, a very impressive accomplishment. Mr. R. Johnson of the Bermuda Biological Station graciously prepared the recent Station 'S' data so they could be included in Figure 8. The Moderate-resolution Imaging Spectroradiometer (MODIS) SST data in Figure 1 were obtained through the Physical Oceanography Distributed Active Archive Center (PO.DAAC), NASA Jet Propulsion Laboratory, Pasadena, CA. http://podaac.jpl.nasa.gov. For the first five years, the Oleander program was funded by NOAA; since then the activity has been supported by grants from the NSF. ONR has also provided very helpful assistance. This paper was completed during a visit by the first author to the University of Stockholm. Last but not least, we thank Dr. W. Sturges and another reviewer for their very helpful comments.

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