Measurement of Volume Transport of the Gulf Stream South of New England[1]

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ABSTRACT

During June 1966, nine velocity measurements were made with neutrally buoyant floats at a depth of about 2500 m along a section crossing the Gulf Stream near 38°N, 69°W. In combination with hydrographic stations bracketing each float trajectory, these measurements yielded a volume transport of $101 \times 10^6 m^3/sec$ for the Stream as of that time and location. The uncertainty that occurs in this figure from measurement error, mixture of time scales from the different types of measurement, and ambiguity in definition of the edges of the Stream is estimated to be 20–30%; this is notably better than figures that can be obtained for this area with arbitrary assumptions of level isobaric surfaces. Although the measured velocities (6–11 cm/sec) are significantly smaller than the few earlier comparable observations (16–17 cm/sec) reported farther downstream by Fuglister (1963), they are large enough to imply a net flow at the bottom in the same direction as the surface current.

Introduction. The extensive program of transport measurement carried out by Richardson and Schmitz (1965) and Schmitz and Richardson (1966) has established with good precision the average volume transport of the Gulr Stream in the Straits of Florida and the magnitude of its typical fluctuations. Downstream from the Straits, however, this primary quantity in any North Atlantic water budget is remarkably uncertain. Iselin (1940) and Fuglister (1963) have published dynamically computed transports based on assumptions, respectively, of zero velocity below 2000 m and zero velocity at the ocean bottom; but Fuglister (1963) has pointed out that a few deep velocity measurements made with neutrally buoyant floats suggest non-zero bottom velocities, implying a possible 50% increment in transport.

Thus, to obtain accurate absolute values of transport, dynamic computations must be tied to velocity measurements closely spaced across the Stream. This method, especially when based on neutrally buoyant floats, seems ineffecient

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and time-consuming, but it appears to be the most reliable one now available for use in the Gulf Stream where it is far distant from the coast.

In any case, the real limitation to useful measurements of the Gulf Stream transport (away from the Straits of Florida) does not lie in the observational techniques but in (i) the inherent ambiguity in the very idea of "Gulf Stream," and (ii), to a degree, in the concept of the volume transport of a current. (Fuglister has previously expatiated on the problems of defining the "Gulf Stream.") Within the Straits of Florida, the term "Gulf Stream" can be used clearly and unequivocally to mean that body of water bounded physically by Florida and the Bahama Bank or Cuba. But after the Stream leaves the Straits, especially downstream from Cape Hatteras where it departs from the Continental Slope, this convenient method of definition is not applicable; here, rather than specifying lateral boundaries, the term "Gulf Stream" serves mainly for pointing from a distance to a particularly large and abrupt change in the depth of the thermocline or to an area of especially intense coherent flow in the North Atlantic subtropical circulation. One can pose arbitrary definitions of Stream edges, but, since the distribution of properties varies in space and time, these are only roughly useful. Some evidence (Volkmann 1962) suggests that, in the slope water between the Stream and the Continental Shelf, the flow is generally westward or southwestward, and a zero-velocity surface separating this flow from the Stream might become a convenient "inshore edge" of the Stream. But there is no obvious feature to use as a distinction between what one calls "Gulf Stream" and "Sargasso Sea." Any measurement of transport of the "Gulf Stream" depends on a particular choice of Stream edges, and it is therefore vitiated by the uncertainty in the definition of the Stream edges; the difficulty is not one of observation but of intrinsic and unavoidable vagueness in physical-geographic nomenclature.

These difficulties are compounded during lateral shifts of the Stream's path where it is not constrained by physical boundaries. Apparently these shifts occur roughly barotropically, implying rotation with depth of the horizontal velocity vector in the Stream itself. Since volume transports of currents are conventionally referred to sections normal to the current, one feels uneasy about the significance of the transport across a single section when the current direction varies markedly with depth; indeed, one may wonder to what extent it makes any sense to speak of a "total volume transport of a current" in these special circumstances.

Despite these discouraging logical limitations, one can, with a certain amount of caution, undoubtedly speak meaningfully of the Gulf Stream transport with a certainty greater than that allowed by measurements to date. In fact, the present uncertainty is so great that even a few reliable transport estimates will be of at least guiding value in working toward the desired quantitative description of the North Atlantic circulation. With this end in mind, a single measurement of the transport of the Gulf Stream was undertaken south of

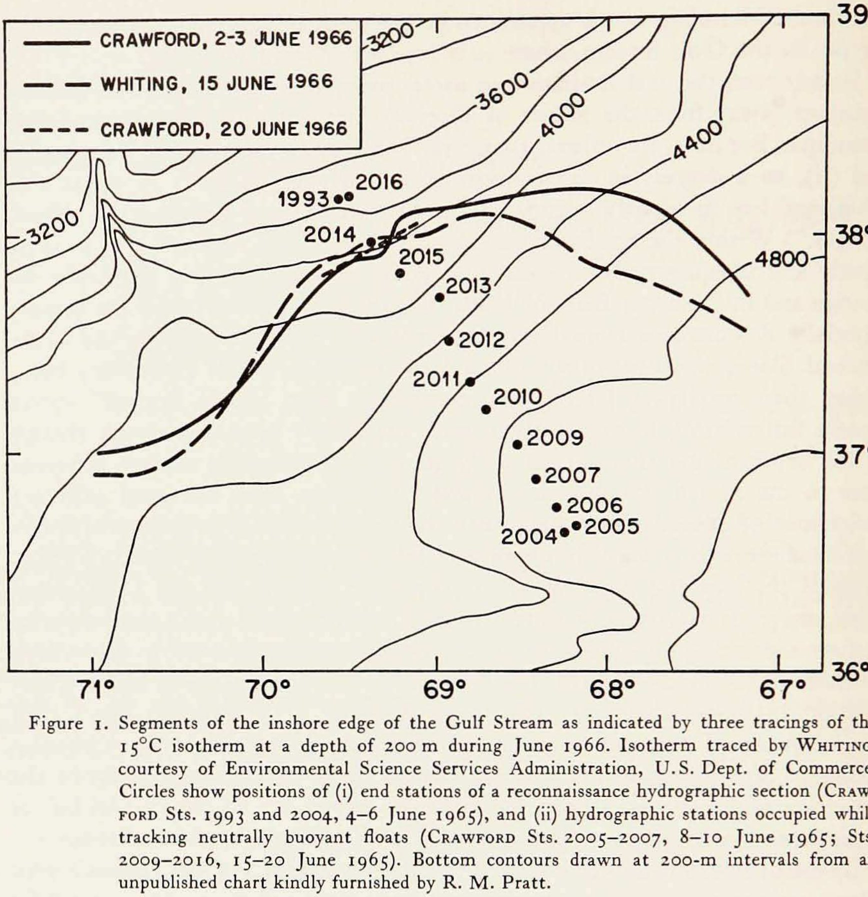


Figure 1. Segments of the inshore edge of the Gulf Stream as indicated by three tracings of the 15°C isotherm at a depth of 200 m during June 1966. Isotherm traced by Whiting, courtesy of Environmental Science Services Administration, U.S. Dept. of Commerce. Circles show positions of (i) end stations of a reconnaissance hydrographic section (Crawford Sts. 1993 and 2004, 4–6 June 1965), and (ii) hydrographic stations occupied while tracking neutrally buoyant floats (Crawford Sts. 2005–2007, 8–10 June 1965; Sts. 2009–2016, 15–20 June 1965). Bottom contours drawn at 200-m intervals from an unpublished chart kindly furnished by R. M. Pratt.

New England by means of combined hydrographic stations and neutrally buoyant floats. The attempt, moderately successful, and instructive concerning its limitations, is reported in this paper.

Observational Program. The observations were made in June 1966 from R/V Crawford. Following an initial survey of the Stream path (Crawford 2–3 June; Fig. 1) with BT's, a reconnaissance hydrographic section was made normal to the Stream, including its full width as indicated by the temperature structure (Sts. 1993–2004; only the end stations of the series are shown in Fig. 1). Starting at the southern end, neutrally buoyant floats were planted in progression across the section at a depth of about 2500 m and at intervals of 10–12 miles. Each float was generally tracked for 2 to 3 days, and as one was abandoned another was launched; thus 3 to 4 floats were always being worked. Concurrently, additional hydrographic stations were occupied so that each float track was bracketed with a pair of stations (Sts. 2005–2007, 2009–2016; Fig. 1). The signals from the floats were weaker than expected, so weak in fact as to frustrate attempted velocity measurements in the slope water at the northern edge of the Stream (between Sts. 2014 and 2016; Fig. 1). The project was interrupted from 10–14 June, following occupation of St. 2007, by the passage of Hurricane Alma. As a final check on the position and direction of the Stream, the cruise concluded with a brief track of the 15° isotherm at 200 m (20 June; Fig. 1).

The observations of the Stream path at three different times (Fig. 1) suggest that, at the section line, the position of the Stream did not shift by more than about five miles nor did its direction vary by more than 10° during the period 2–20 June. The implied average horizontal velocity components normal to the isotherms would then be less than 1 cm/sec, leading to insignificant ambiguity in the volume transport measurement due to lateral shifting.

The general use of neutrally buoyant floats in making direct current measurements has been described by Swallow (1955). The model used on this occasion houses the batteries and electronics in a 10-inch glass sphere and operates at 4 to 6 khz. A more accurate timing method permits recording of the signal on a facsimile-type recorder, although the two-beam oscilloscope was still used for fine resolution. By towing a three-hydrophone array, the ship was maneuvered directly above the float, so that taking a fix on a float was reduced simply to getting a fix on the ship. (A full report on this modified system is being prepared by D. C. Webb and G. H. Volkmann.) Loran A was used for navigation throughout this cruise; three stations were read for every fix.

Although all salinities were determined by the conductivity method, it was not feasible on this occasion to run the analyses at sea. Unfortunately, an additional 2 to 3 months elapsed before the water was analyzed at Woods Hole, and apparently this time interval was long enough for sufficient evaporation to take place through minute imperfections in certain bottle caps to degrade some of the samples; 13% of the salinities had to be rejected as too high, either because they differed by more than two standard-deviation units from the mean temperature-salinity curve for the region (Worthington and Metcalf 1961), or because they were associated with apparent static instability. It would probably be dangerous to regard the accuracy of salinities retained as better than ± 0.01 ‰.

Profiles of Temperature, Salinity, and Specific Volume Anomaly. The cross-stream slope in isotherms, isohalines, and isanosteres that are characteristic of the Stream is clearly evident in the profiles of temperature, salinity, and specific-volume anomaly for Sts. 2005–2007, 2009–2016 (Figs. 2–4). Although no particular leveling of isopleths is apparent between Sts. 2014 and 2016 to

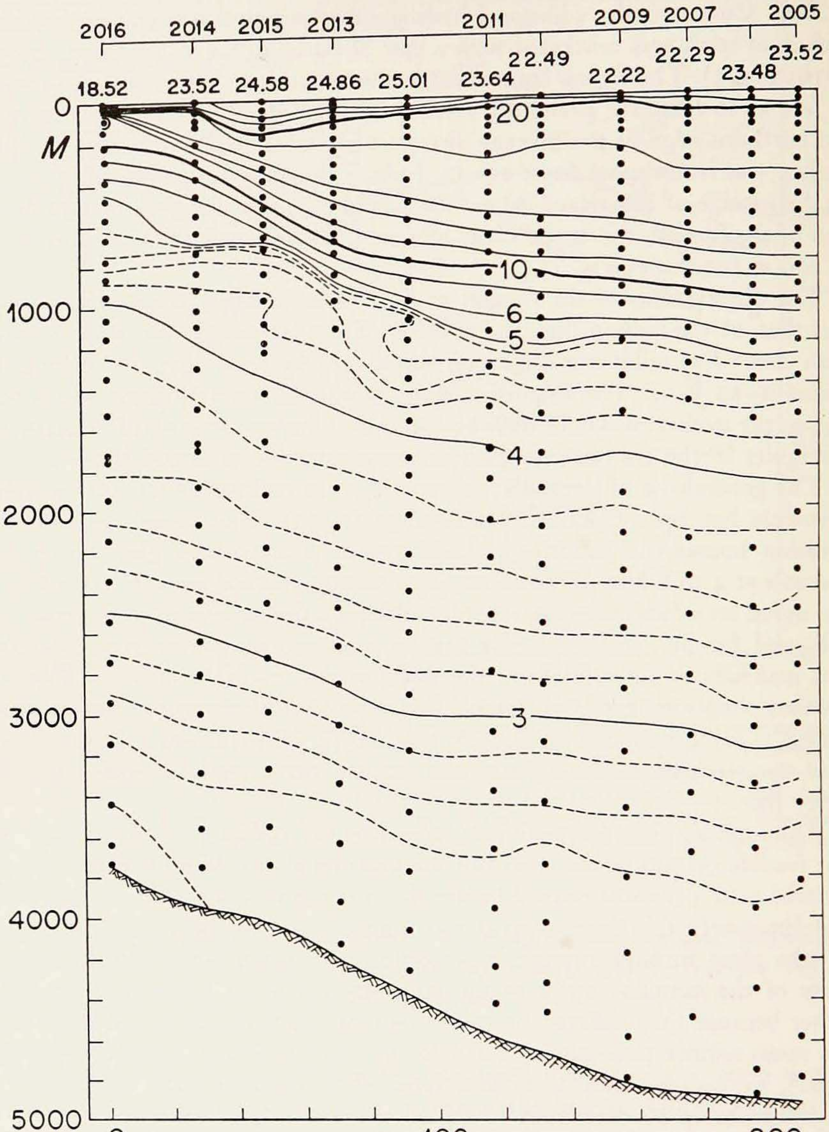


Figure 2. Profile of temperature (°C) across the Gulf Stream, CRAWFORD Sts. 2005–2007, 2009–2016. Isotherms are drawn at intervals of 2°C above 6°C and at intervals of 0.2°C below 5°C. Dots indicate positions of observations; see Fig. 1 for location and dates of stations. Depths in meters; bottom topography interpolated from soundings made at stations. A scale of distance (km) along the section is given at the bottom of the profile, with station numbers and surface values along the top; the indicated station spacing is based on positions projected onto an average straight line for the section.

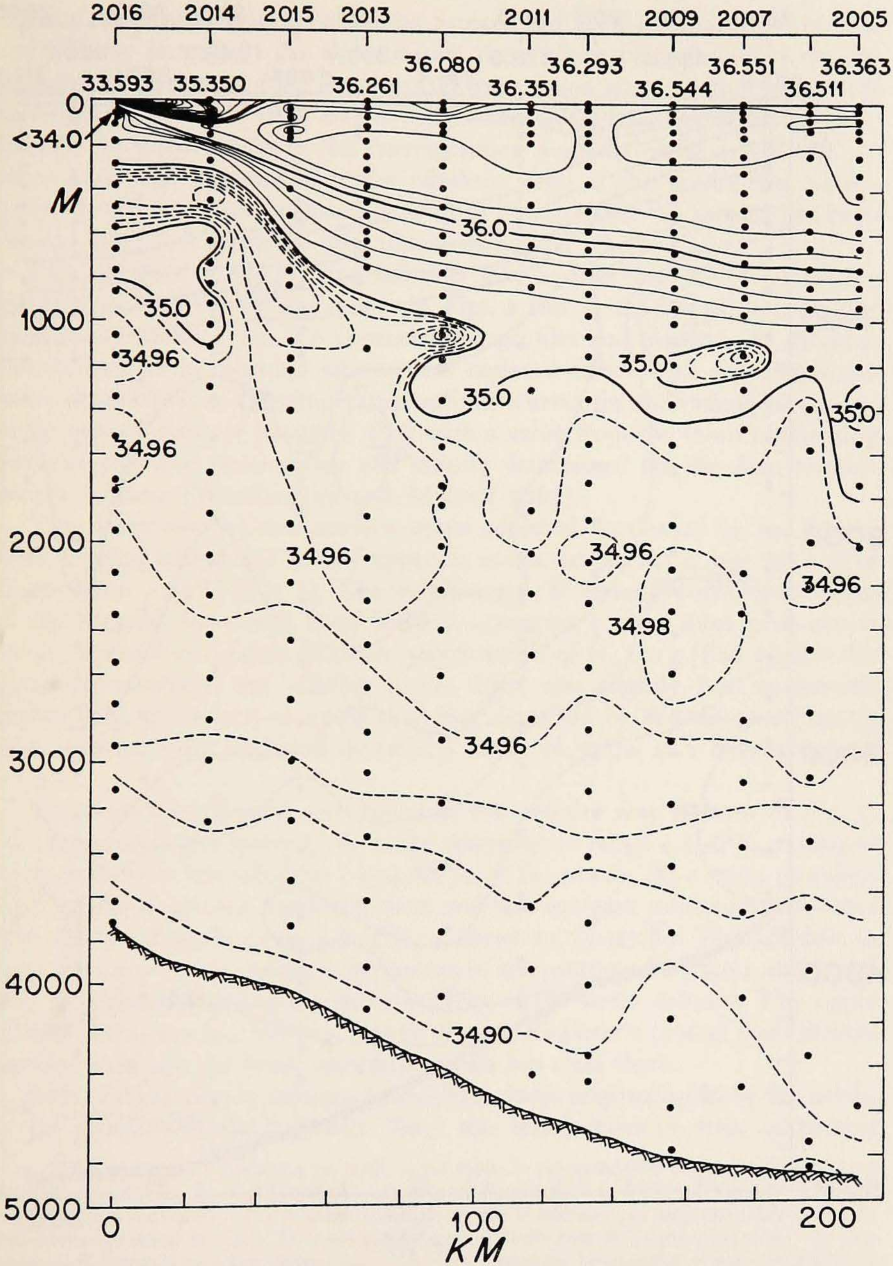


Figure 3. Profile of salinity (‰) across the Gulf Stream, CRAWFORD Sts. 2005–2007, 2009–2016. Isohalines are drawn at intervals of 0.2‰ above 35.00‰, at intervals of 0.02‰ below 35.00‰. See caption for Fig. 2.

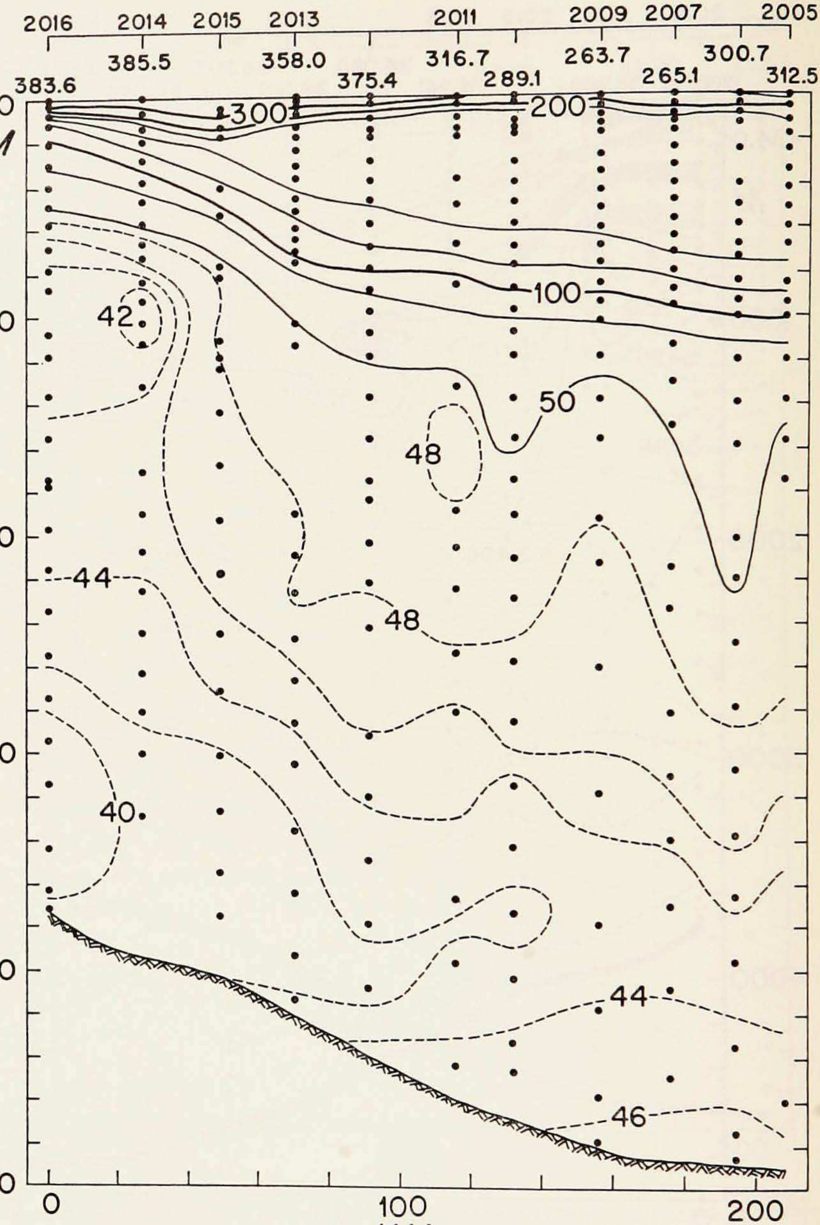


Figure 4. Profile of anomaly of specific volume (cl/t) across the Gulf Stream, CRAWFORD Sts. 2005–2007, 2009–2016. Isanosteres are drawn at intervals of 100 cl/t above 200 cl/t, 25 cl/t between 200 and 50 cl/t, and 2 cl/t below 50 cl/t. Dots indicate positions of paired observations of temperature and salinity. See caption for Fig. 2.

indicate that the inshore edge of the Stream had been crossed, St. 2016 was undoubtedly located in the slope water; the inshore pair of stations on the reconnaissance section (Sts. 1993–2004), mentioned above, which were only half as far apart as Sts. 2014 and 2016, showed nearly level isopleths at approximately the same depths as the corresponding isopleths on St. 2016. On the other hand, the isopleths near the off-shore edge of the profile (Sts. 2005–2007, 2009–2010) are sufficiently level to indicate that the Stream had been crossed, though it would be difficult to decide just where.

The locations of the rejected salinities may be seen in detail by comparing the positions of observations shown in Figs. 2 and 3; the loss of data was the most severe at St. 2005. To construct the profiles and perform the dynamic calculations, every rejected salinity was replaced either with an interpolated value obtained from the temperature-salinity curves for the individual stations or (at temperatures of less than 3°C) with a value from the mean relationship between potential temperature and salinity determined for the deep western North Atlantic (Worthington and Metcalf 1961).

The warm core of near-surface water advected northward by the Stream from tropical latitudes is readily apparent at Sts. 2012, 2013, and 2015 as of temperature >24°C (Fig. 2). The well-known 18° water (Worthington 1959) of the Sargasso Sea—and Gulf Stream—occupies a layer some 300–400 m thick, centered at a depth of about 300 m south of St. 2013 (Fig. 2). On this particular occasion, the salinity in the layer was slightly but consistently higher (Fig. 3) by 0.05–0.15‰ than that described by Worthington (1959), who reported little variation in salinity from 36.50‰ at a temperature of 17.9°C.

At intermediate depths, two layers of low salinity may be seen in Fig. 3: (i) a thin and rather spotty layer in the temperature range 4.5–6°C, σ_t around 27.6, with very low salinities (<34.88‰ at St. 2012); (ii) a more prevalent layer of much greater vertical extent and less extreme minimum associated with the temperature range 3.6–4°C, σ_t about 27.75–27.80. The salinities in layer (i) are so low as to allow inversions in temperature (Sts. 2012 and 2015, Fig. 2) without violating the static stability of the water column. The upper salinity minimum is a striking feature also on Fuglister's (1963) Gulf Stream profiles, although the lower minimum seems less clear there.

Both of these salinity minima derive from water originating far to the north, in the environs of the Labrador Sea;[2] this water flows or seeps southward

2. The nomenclature pertaining to these water types is extraordinarily confused. Wüst (1935) identified a water mass in the Labrador Sea area that is characterized by a salinity minimum at depths of 500–1000 m and a rather variable temperature and σ_t; this is obviously the shallower of the two intermediate-depth minima in Fig. 3. He preferred the name "North Atlantic intermediate water" for this water mass, although he acknowledged use of the term "subarctic intermediate water," which is the name adopted by Dietrich (1963). Wüst did not discuss the deeper of the two salinity minima *per se*, but rather an intermediate-depth maximum in dissolved-oxygen concentration, which appears to be a diagnostic of the same water mass (σ_t about 27.80, depth 2000–2500 m in the Sargasso Sea and 200 m in the slope water and Labrador Sea): this he labeled "intermediate North Atlantic deep water." Smith

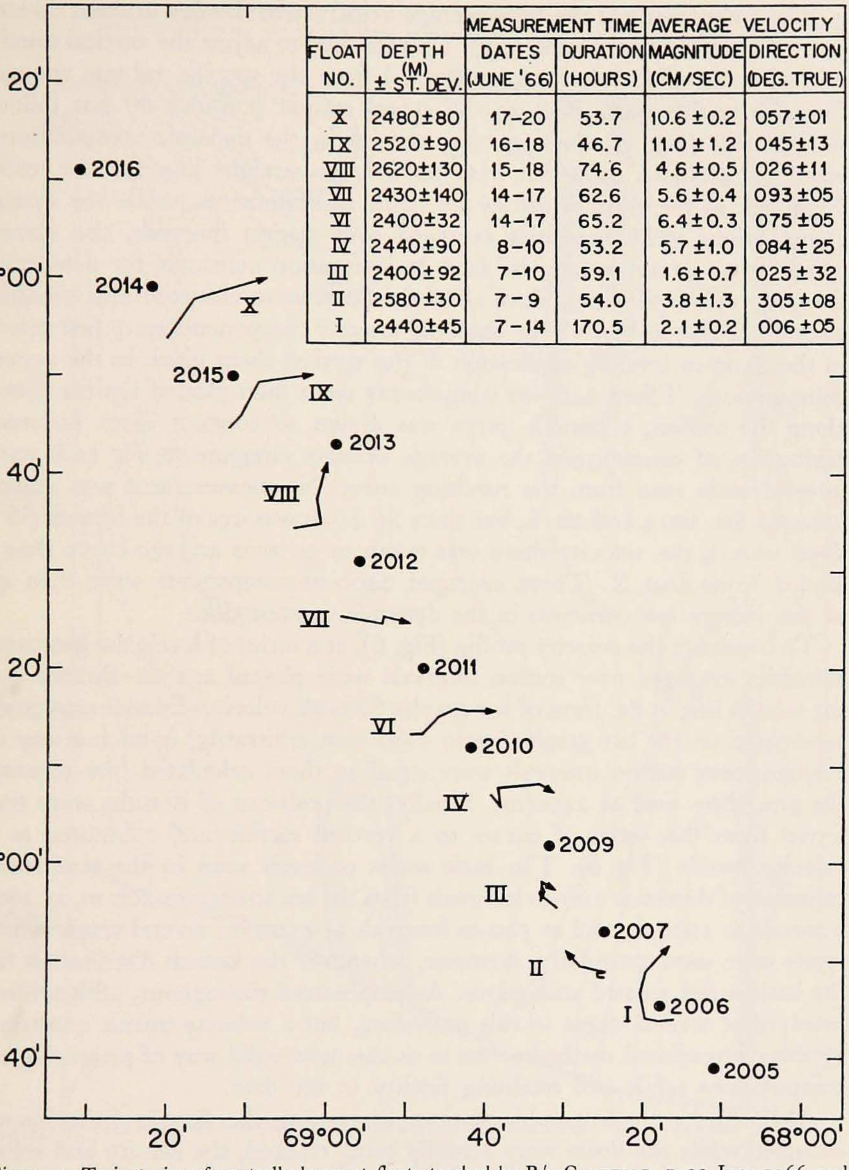
around the Grand Banks into the slope-water area. Its vivid presence here in the Gulf Stream, even out to the edge of the Stream adjacent to the Sargasso Sea (Sts. 2007 and 2009; Fig. 3), must imply a very considerable inflow to the Stream from the slope water; probably this is part of the great increase in volume transport of the Stream that seems to occur near and somewhat downstream from Cape Hatteras.

The pronounced layer of relative minimum in specific volume anomaly (δ) at 3000–4000 m (Fig. 4) does not imply static instability. At these depths the vertical variations in temperature and salinity are small and are in such proportion that thermosteric anomaly (or σ_t) is virtually constant with depth; however, in the conventional formula for δ, the term that expresses the cross-dependence of temperature and pressure, is, at great depth, of the same magnitude as the thermosteric anomaly; and this quantity of course increases with depth, thus giving the deep inversion in δ.

Velocities. The trajectories of the nine floats that provide usable information fall into three groups (Fig. 5). The three southernmost floats (Nos. I–III) moved erratically, with no mutual coherence; the individual tracks show little directional consistency with time and are therefore associated with small average velocities (Table in Fig. 5). In contrast, the three central floats (Nos. IV, VI, VII) had trajectories very similar to one another, showed only small directional variation in time, and moved predominantly in the direction of the surface Stream but a little southward of the downstream normal (direction 057°T) to the section line. The three northernmost floats (Nos. VIII–X) showed somewhat similar coherence and directional consistency but had velocity components northward of the normal to the section. The rather anomalous behavior of No. VIII should be noted. Of the nine floats, Nos. IX and X moved the most rapidly, as might be expected, since they lay beneath the swiftest part of the current in the near-surface water. (This grouping does not appear to have resulted from the interruption in work caused by the passage of Hurricane Alma, because the work was broken off only after the observations on Nos. I–IV were completed.)

The general shapes of the trajectories suggest that floats IV and VI–X were in the Gulf Stream, but that Nos. I–III were not, despite the occurrence of water from the slope-water area at St. 2007. We therefore regard the Stream as bounded by Sts. 2009 and 2016.

et al. (1937), while recognizing the shallower minimum and accepting Wüst's name for it, called the deeper salinity minimum "intermediate water of the Labrador Sea." The description by Sverdrup et al. (1942: 670) of intermediate water originating in the North Atlantic is very cursory; they called attention to one minimum and named it "Arctic Intermediate Water," but it is not fully clear which of the two minima this is. After examining the two stations cited by them as exemplifying this water mass, we believe that they were actually referring to the deeper minimum; Barrett (1965) made the same interpretation and followed their nomenclature. Recently, Lee and Ellett (1967) called this water mass simply "Labrador Sea Water."



FLOAT NO.	DEPTH (M) ± ST. DEV.	MEASUREMENT TIME DATES (JUNE '66)	MEASUREMENT TIME DURATION (HOURS)	AVERAGE VELOCITY MAGNITUDE (CM/SEC)	AVERAGE VELOCITY DIRECTION (DEG. TRUE)
X	2480±80	17–20	53.7	10.6 ± 0.2	057±01
IX	2520±90	16–18	46.7	11.0 ± 1.2	045±13
VIII	2620±130	15–18	74.6	4.6 ± 0.5	026±11
VII	2430±140	14–17	62.6	5.6 ± 0.4	093 ±05
VI	2400±32	14–17	65.2	6.4 ± 0.3	075 ±05
IV	2440±90	8–10	53.2	5.7 ± 1.0	084 ± 25
III	2400±92	7–10	59.1	1.6 ± 0.7	025 ± 32
II	2580±30	7–9	54.0	3.8 ± 1.3	305 ±08
I	2440±45	7–14	170.5	2.1 ± 0.2	006 ±05

Figure 5. Trajectories of neutrally buoyant floats tracked by R/v CRAWFORD, 7–20 June 1966, and positions of hydrographic stations (2005–2007, 2009–2016) occupied concurrently. Time and duration of observations on floats I–IV, VI–X, and depths and measured average velocities of floats are summarized in the Table. Depths were determined from direct and bottom-reflected signals from floats when they were directly beneath the ship; error estimates refer to standard deviations of repeated determinations. The depth used for float X is the average of all others, because an acoustic determination was not possible. Error estimates for average velocities refer to "triangles of error" in initial and final fixes on floats as determined by three-channel Loran A fixes on ship.

The components of the nine average velocities in the downstream direction normal to the station section (057°T) were used to adjust the vertical distribution of geostrophic velocity as computed from the specific volume anomalies for each station pair. Because the actual station positions do not follow a straight line (Fig. 1), their spacing as used in the dynamic computations is based on positions projected onto an average straight line for the section. Inasmuch as the float velocities are point measurements, while the dynamic computations yield quantities averaged over station intervals, the observed components cannot strictly be used as integration constants for determining the geostrophic velocities. Since all of the floats were clustered near a depth of 2400 m (Table in Fig. 5), the observed velocity components were first referred to the 2400-m level by application of the vertical shear given in the dynamic computations. These 2400-m components were then plotted against distance along the section, a smooth curve was drawn to connect them (somewhat arbitrarily, of course), and the average velocity components for each station interval were read from the resulting curve. No measurement was available between Sts. 2014 and 2016, but since St. 2016 was out of the Stream (in the slope water), the velocity there was taken to be zero and the curve thus extended from float X. These averaged 2400-m components were then used as the integration constants in the dynamic computations.

To construct the velocity profile (Fig. 6), at a series of levels the geostrophic velocities averaged over station intervals were plotted against distance along the section line in the form of bar graphs. Smooth velocity-distance curves were superposed on the bar graphs (again somewhat arbitrarily) in such a way that averages over station intervals were equal to those calculated (the reverse of the procedure used at 2400 m). Finally, the positions of isotachs were transferred from this series of curves to a vertical section and contoured as the velocity profile (Fig. 6). The basic series of levels used in the construction consisted of depths at 100-m intervals from the sea surface to 200 m, at 200-m intervals to 1000 m, and at 500-m intervals to 4500 m; several supplementary levels were used irregularly, however, whenever the isotach distribution from the basic series seemed ambiguous. As emphasized throughout, arbitrariness is involved at several stages in this procedure, but a velocity profile constructed by this conventional method seems to us the most vidid way of presenting such computations while still retaining fidelity to the data.

Although the hydrographic stations bracketing the float trajectories were occupied while the floats were actually being tracked, the density and velocity observations represent very different time-scales of measurement; since they were spread out over a two-week time period as well, it is difficult to know in how much detail a velocity profile such as that in Fig. 6 describes real features (or some time-average of real features) in the ocean. Internal waves, for example, notoriously distort geostrophic velocity fields as derived from dynamic computations, and one cannot be sure what relation the measure-

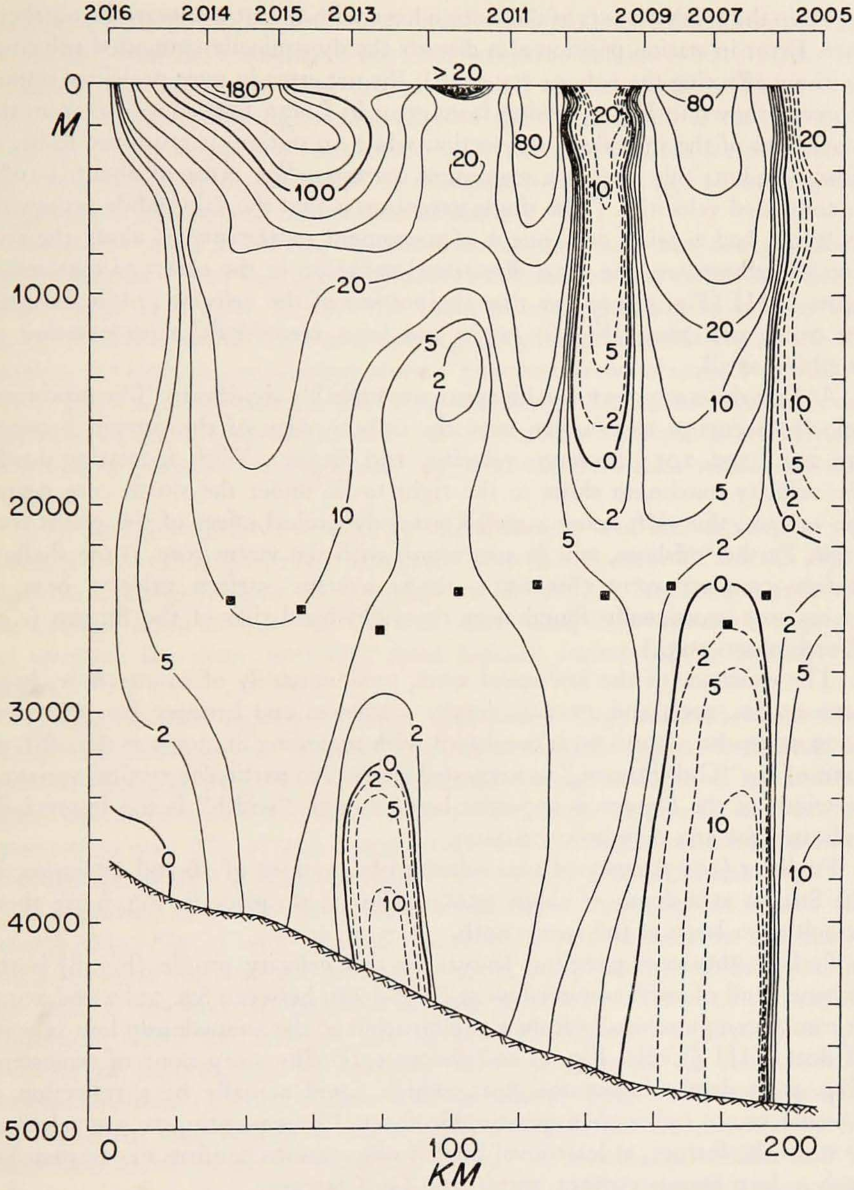


Figure 6. Profile of geostrophic velocity (cm/sec) across the Gulf Stream, constructed from observed velocities of neutrally buoyant floats I–IV, VI–X (Fig. 5) and measured specific-volume anomaly at CRAWFORD Sts. 2005–2007, 2009–2016, as described in text. Isotachs are drawn at intervals of 20 cm/sec, with the isotachs of 2, 5, and 10 cm/sec also included. Dashed isotachs indicate flow counter to the Gulf Stream. Solid squares show positions of neutrally buoyant floats I–IV, VI–X on profile; see Fig. 5 for individual identification. See caption for Fig. 2.

ments in the southern part of the section have with the later ones in the northern part. Error in station position also distorts the dynamically computed velocities (without affecting the volume transport); the net error in most projected station intervals shown in Fig. 5, arising from error in Loran fixes as well as from the movement of the ship along the section while on station, is estimated to be, at most, ± 2 km; this implies a maximum corresponding error of about $\pm 10\%$ in computed velocities. (The single exception is that the ship, while occupying St. 2006, had a 7-km component of movement southeastward along the section.) Furthermore, the great directional variation in the observed motions of floats I–III (Fig. 5) suggests that the portion of the velocity profile between Sts. 2005 and 2009 (Fig. 6) might not be a meaningful representation of anything at all.

At least the coarse features, however, are probably significant. The maximum velocity occurs at the surface near the inshore edge of the Stream between Sts. 2014 and 2015 (average velocity, 179 cm/sec). With increasing depth, the velocity maximum shifts to the right to lie under the warm core nearer the surface, this shift being a well-known dynamical effect of the warm core itself. Farther offshore, still in association with the warm core, is the shallow surface countercurrent (Sts. 2011–2012; average surface velocity here of 11 cm/sec) occasionally found near the right-hand side of the Stream (*e.g.*, Worthington 1954).

The existence of the low-speed zone, predominantly of counterflow, lying between Sts. 2009 and 2010 at depths <2200 m and between Sts. 2007 and 2009 at depths >2200 m, is consistent with regarding St. 2009 as the offshore limit of the "Gulf Stream," as suggested above. No particular vertical variation in width of the Stream is apparent here, though "width" is too imprecisely indicated for any definite conclusion.

Fuglister (1963) reported two velocity observations of 16 and 17 cm/sec in the Stream at a depth of about 2500 m, but at no place in Fig. 6 are there velocities so large at the same depth.

Perhaps the most puzzling feature in this velocity profile (Fig. 6) is the narrow band of swift counterflow at great depth between Sts. 2012 and 2013. It results computationally from a combination of the anomalously low velocity of float VIII (Table, Fig. 5) and the exceptionally steep slope of isanosteres (Fig. 4) at depths below the float, which could actually be a reflection of internal waves rather than geostrophic shear. Consequently, we are reluctant to stress the feature, at least until further observations confirm the existence of such a deep countercurrent within the Gulf Stream.

Transports. The total volume transport across the entire section (Sts. 2005–2007, 2009–2016), fitted to the measured float velocities by the procedure described, is $117 \times 10^6 m^3/sec$. The Gulf Stream itself, however, should probably be limited to Sts. 2009–2016; the net transport across this part of the section is $100 \times 10^6 m^3/sec$, and if only the flow moving with the surface Stream is counted, then the transport of the Gulf Stream so defined is $101 \times 10^6 m^3/sec$.

Error from various sources can be assigned to this number. There is no clear definition of the right-hand edge of the Stream, and if the eastward-moving flow between Sts. 2005–2009 were added to the $101 \times 10^6 m^3/sec$ for the remaining stations, the resulting transport would be $123 \times 10^6 m^3/sec$, thus suggesting a definitional imprecision of about 20%.

Since the water below the thermocline moves slowly but occupies about three-quarters of the full depth, transport calculations are rather sensitive to small errors in density. We do not claim a better accuracy in salinity than ± 0.01 ‰; a constant error of such magnitude in all observations at one end station, for example, would introduce a transport error of some $8 \times 10^6 n^3/sec$ for a 4000-m water column with zero velocity at the top or bottom. A distribution of undetected salinity error so systematic with depth is unlikely, but it indicates the order of magnitude of transport errors with which one must reckon.

The navigational error in velocity measurements is of less significance: the average error of ± 0.6 cm/sec associated with Loran error (Table in Fig. 5) leads to a transport error of only $\pm 4 \times 10^6 m/^3/sec$ for Sts. 2009–2016, with an average depth of 4400 m. On the other hand, it is impossible at present to estimate the error (probably more serious) introduced by mixing time scales.

Compared with regions of slight quasisteady flow, internal waves cause much less uncertainty in transport computations when applied to features like the Gulf Stream, where the net cross-stream change in depth of isanosteres is great compared with wave amplitudes. Even a large amplitude such as 50 m amounts to only 6% of such characteristic depth changes in isanosteres in the thermocline as 800 m (Fig. 4), with a corresponding error in transport.

Comparison of results from the reconnaissance hydrographic section (Sts. 1993–2004, Fig. 1) with those of the repeated section made in conjunction with the float observations (Sts. 2005–2007, 2009–2016) affords a practical instance of the general uncertainties associated with dynamically computed transports. With the arbitrary assumption for comparative purposes of zero velocity at 3000 m, the net transport above 3000 m on the reconnaissance section would be $111 \times 10^6 m^3/sec$; with the same assumption, the corresponding figure for the later section would be $92 \times 10^6 m^3/sec$. Since approximately four days intervened between completion of the reconnaissance section and beginning the second section—about one-third of the time required to complete the second section (12 days), the large difference between the two computed transports across essentially the same section must imply an uncertainty at least as big in the transport number computed from the paired stations and float observations. Thus a combination of the various errors arising from explicit measurement error, inappropriate application of the geostrophic approximation, mixture of time scales, and inherent ambiguity in language suggests a significance level of about 20–30% for the volume transport number of $101 \times 10^6 m^3/sec$.

A by-product of calculations such as these is an estimate of the bottom velocities in the Gulf Stream, which is necessary for any assessment of the effect of bottom topography on the path of the Stream (Warren 1963). With the dynamic computations for the individual station pairs tied to the float velocities as described above, the cross-stream integral of the bottom velocity between Sts. 2009 and 2016 is $2.2 \times 10^7 cm^2/sec$ in the direction of the surface current, implying a cross-stream averaged bottom velocity of about 1.5 cm/sec. Although the error in float velocities (about ± 0.6 cm/sec on average) leads to wholly insignificant uncertainties in total volume transport, it is obviously much more serious in estimating bottom velocities. Of greater concern, however, must be the representativeness of the velocity measurements and their strict application to such calculations. The strong counterflow at great depth between Sts. 2012 and 2013 (Fig. 6), which is required by the anomalous, possibly small-scale transient behavior of Float VIII, for example, makes a large negative contribution of $1.7 \times 10^7 cm^2/sec$ to the cross-stream integral. Perhaps, therefore, this estimate should not be supposed accurate to better than $\pm 50\%$.

Nevertheless, the bottom velocities calculated here are significantly and markedly smaller than the values of about 8 cm/sec implied by the several earlier deep-current measurements downstream from this section listed by Fuglister (1963), and smaller also than the single measurement of bottom velocity (10 cm/sec) made upstream from the section by Knauss (1965). It is quite impossible, of course, to tell whether this difference is one primarily of time or of location.

A bottom transport per unit depth so low as $2.2 \times 10^7 cm^2/sec$ is not compatible with an explanation of the local meander pattern as a combination of topographic and Rossby waves (Warren 1963). The magnitude of the depth gradient near this section is about 8×10^{-3}, and its direction, about 140°; the bottom transport per unit depth and the net volume transport between Sts. 2009 and 2016 are given above, and the momentum transport (per unit mass) works out to be about $6.1 \times 10^{15} cm^4/sec^2$; the current path appears to inflect near 37°15′N, 70°15′W (Fig. 1), where the current direction is about 060°T; and in these latitudes $f = 0.9 \times 10^{-4} sec^{-1}$, $\beta = 1.8 \times 10^{-13} cm^{-1} sec^{-1}$. With these data, the formulas given by Warren (1963) call for a meander amplitude of 105 km and a half-wavelength of 340 km. The observed anticyclonic meander (Fig. 1), however, has an amplitude of roughly 75 km and a half-wavelength of some 270 km, both values being only about three-fourths of those calculated. The uncertainty in the actual bottom transport per unit depth might just be large enough to tolerate this discrepancy, but lengthy speculation would be pointless.

Concluding Remarks. It may be concluded that measurement of volume transport in the Gulf Stream by the method described here is useful if one is prepared to accept uncertainties of 20–30% in the individual measurements. Given the greater uncertainty in arbitrary assumptions of level isobaric surfaces and the vagueness in meaning of "Gulf Stream transport" downstream from Cape Hatteras, such error as this seems tolerable at present. Obviously, however, one cannot hope to detect variations (either in time or space) of lesser magnitude.

The other noteworthy result derived from this cruise is the detailed direct measurement of flow in the deep water, which gave velocities that are lower by a factor of at least two than the few comparable measurements described by Fuglister (1963). Only extensive future observations will be able to determine whether such a difference is more typically one of time or of space.

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