PAPERS IN PHYSICAL OCEANOGRAPHY AND METEOROLOGY

PUBLISHED BY

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

AND

WOODS HOLE OCEANOGRAPHIC INSTITUTION

VOL. VIII, NO. 1

PRELIMINARY REPORT ON LONG-PERIOD VARIATIONS IN THE TRANSPORT OF THE GULF STREAM SYSTEM

BY

C. O'D. ISELIN

Contribution No. 261 from the Woods Hole Oceanographic Institution

CAMBRIDGE AND WOODS HOLE, MASSACHUSETTS

July, 1940

CONTENTS

INTRODUCTION

In 1937 the Woods Hole Oceanographic Institution and the Bermuda Biological Station for Research agreed to cooperate on a five year program of observations. This was designed to throw light on the general problem of long-period fluctuations in the transport of the Gulf Stream, and their possible significance for fisheries research and for meteorology. The investigation originated at the suggestion of the Bermuda Oceanographic Committee of the Royal Society of London, which also arranged for a generous grant of money to enable the Bermuda laboratory to undertake part of the field work.

A summary of the ideas on which the original program was based has already been published (Iselin, 1938a) and more recently two short papers (Iselin, 1938b, 1939) have emphasized certain aspects of the theoretical considerations on which it rests. Although the primary objective has been a study of long-period trends in the strength of the Gulf Stream and only two and a half years have elapsed since the work at sea began, nevertheless it now seems desirable to set forth some of the preliminary results and to discuss further the underlying assumptions. It is hoped that in this way we can gain the benefits of criticism and thus be more wisely guided during the remainder of the five years for which continued field work is now planned.

In preparing this report I have had the assistance and advice of the staff of the Woods Hole Oceanographic Institution. I am especially indebted to Dr. Sidney C. T. Hsiao who prepared the final drawing of the diagrams and carried through the necessary computations. In addition, I have worked in close cooperation with Dr. E. F. Thompson of the Bermuda Biological Station and Mr. H. B. Hachey of the Atlantic Biological Station. It is most gratifying to have taken part in an investigation with so many willing and helpful participants.

OBSERVATIONAL APPROACH

Since it is not practical to measure directly the transport of ocean currents and since it was necessary to work out an observational program for which a single ship would suffice at the outset of this investigation we were forced to make certain assumptions.

In the first place, of course, it was tentatively assumed that the Gulf Stream varied in transport over periods of several months or more, and that the changes were large enough so that they could be measured. There has long been a general belief that the major ocean currents do fluctuate in strength and that these changes are from time to time transmitted to the fishing grounds, being partly responsible for the variations in the stocks of fish. However, there is no general agreement as to the mechanism whereby these fluctuations of the offshore currents reach the coastal waters (Iselin, 1938b, 1939). In addition, it is usually thought that changes in the oceanic circulation could somewhat modify the weather in certain coastal regions where the winds blow quite regularly from the sea. But there exist very few observations[1] which definitely show that significant long-period fluctuations of any of the major oceanic currents have actually taken place. It is true that in 1931–32 five serial sections, crossing the Gulf Stream off Chesapeake Bay (Iselin, 1936), did indicate a gradual decrease in the cross-current density gradient

[1] A notable exception is the annual surveys which the U. S. Coast Guard has been making since 1926 in the Labrador Current and West Greenland Current.

for this period. However, the series was much too short to rule out the possibility that these were only random fluctuations or that they resulted from the inaccuracies of the methods used. On the whole it does not seem unreasonable to suppose that such a current as the Gulf Stream does vary sufficiently in transport so that the changes can be followed by some indirect observational technique.

In the second place, it was assumed that Bjerknes' circulation theorem may be applied in the usual manner to the case of the Gulf Stream as it flows between Bermuda and New England. This means that it was taken for granted that the pressure distribution can be computed from the observed mass distribution on the basis that pressure gradients vanish along a certain deep level surface, and that the current is such as to have a deflecting force which just balances the resulting pressure distribution. It was also assumed that this current is so powerful that fluctuations in its transport, even should they amount to only a small percentage of the average flow, would show up with some significance by this method. These assumptions are not usually questioned when applied to the Gulf Stream, but it is worth while to point out that in making them we automatically rather closely define the current and also to some extent fix our methods of procedure.

In relying on the circulation theorem one is not apt to question, for example, the idea that in the Gulf Stream the velocity decreases with depth in the upper 2000 meters. In the present investigation it is not necessary to justify the selection of any particular lower limit for the current, for we are not as much interested in its exact volume as in changes in volume. These will be shown up as long as the same base plane is used for each set of observations and as long as within the current and above this level there does not occur a prolonged reversal in the direction of flow.

Another requirement of the use of the circulation theorem is that sections crossing the current and completed within two or three days must be treated as though the stations were occupied simultaneously. Therefore, short-period variations in transport are ruled out. In justification it should be noted that the momentum of the current is huge in relation to the local forces which might in a few days time disturb its transport. Provided some procedure can be found to eliminate most of the effect of short-period internal waves on the dynamic computations, it seems probable that the current does flow sufficiently steadily so that only very small changes in the total cross-current density gradient can occur during the time required to sail across it and to complete the subsurface observations.

This simplification might not be allowable if in this investigation we were concerned with changes in velocity, for the maximum velocity of the current no doubt varies much more rapidly than does its transport. However, as will be explained below, it seems likely that volume fluctuations alone could have such far reaching consequences as to influence the fisheries or the coastal climate.

In short, by relying on the circulation theorem, it is to be expected that by periodically observing the cross-current density gradients in the upper 200 meters or more, the relative changes in the strength of the current can be determined. It may not be possible to compute absolute volumes from each series of observations, but the transport values obtained should accurately demonstrate the degree of variability. There remains only the question of how often and in how much detail the routine subsurface observations must be made.

This was the first practical problem which we undertook to solve, beginning in June 1937. Meanwhile valuable evidence has been gained concerning the internal structure of the Gulf Stream and 15 series of subsurface observations have been obtained to demonstrate the variability of its transport.

At the outset of the investigation it was hoped, as suggested by Jakhelln (1936), that the routine observations could eventually be reduced to work at two fixed stations, one on either side of the current, and that if a procedure could be devised to eliminate most of the effect of short-period internal waves, the repeated occupation of these two stations would give a reliable picture of the gradual changes in transport. Within eight months after the field work began it was evident that this hope was unfounded, for two factors had been underestimated. First, the position of the axis of the Gulf Stream off New England fluctuates laterally more than had been anticipated and secondly, powerful eddies from time to time upset the internal distribution of density on both sides of the current. Thus either one of a pair of fixed stations, which yield a good value for the transport on one occasion, may at another time lie either within the current or within an eddy. Unfortunately, as will be shown below, the eddies were particularly numerous during the second half of 1937 and the first part of 1938. For this reason, it has seemed unwise to derive transport values from the observations secured during the first year of the program, except on the three occasions when complete hydrographic sections were obtained.

It was of course recognized in 1937 that, even if we were able to show up gradual changes in the strength of the Gulf Stream by the use of the circulation theorem, unless a research vessel could be employed in this work on a full time basis, the number of transport determinations would probably be too small to distinguish clearly between the seasonal cycle and variations having a longer trend. Therefore, it was obviously highly desirable to develop some indirect method which would give a more continuous measure of the trend of the current's strength. If this method could be an inexpensive one, the investigation might perhaps be continued indefinitely. Thus the second practical difficulty which had to be faced in 1937 was that unless the reliability of some secondary method could be established during the relatively short period when subsurface observations were being collected more or less regularly, the investigation might terminate before the oceanographic sections had uncovered any marked change.

Fortunately the tide gauge method, suggested by Sandström (1903), Dietrich (1937) and Montgomery (1938), provides a continuous and inexpensive record of the variations in the cross-current density gradient, if it is assumed that the average surface velocity varies with the total transport of the current. It also allows the current to be studied at several different points along its path between the Straits of Florida and Cape Hatteras. Therefore, in February 1938, through the cooperation of the U. S. Coast and Geodetic Survey, a tide gauge was set up at Cat Cay, a small island directly across the Straits of Florida from Miami. After some difficulty in finding a good exposure,[2] a second instrument was installed at Bermuda. Due to various delays the new Bermuda records have only been coming in regularly since May 1939. As will be shown below, these two instruments on the right hand side of the current now seem to be of secondary importance for the purposes of this report as compared to the primary tidal stations of the U. S. Coast and Geodetic Survey along the southeastern coast of the United States. However, this is a tentative conclusion and it will be wise to study further the records from Cat Cay and Bermuda during the next few years.

Another approach to the problem of long-period Gulf Stream fluctuations is provided

[2] During the years 1833–1843 and 1933–1936 a tide gauge was in operation at Bermuda, but in neither case was the location of the instrument free from criticism.

by the thermograph records collected by passenger steamers on the New York to Bermuda and the Boston to Bermuda runs. In analyzing these thermograms it has again been necessary to make certain assumptions and these will be discussed below (page 28).

The chief justification for publishing this preliminary report is that to date all three methods yield consistent results. While two and a half years is much too short a time in which to establish a long-period trend to the Gulf Stream, this agreement between the various approaches to the problem seems to justify the theoretical considerations on which they are based.

TEMPERATURE AND SALINITY PROFILES

The routine oceanographic sections across the Gulf Stream were made on a line between Montauk Point, on the eastern extremity of Long Island, and Bermuda. This particular section (Fig. 1) was chosen because the inshore stations would be accessible from Woods Hole, and because previous experience indicated that off Montauk Point the distance between the Gulf Stream and the edge of the continental shelf was relatively great.

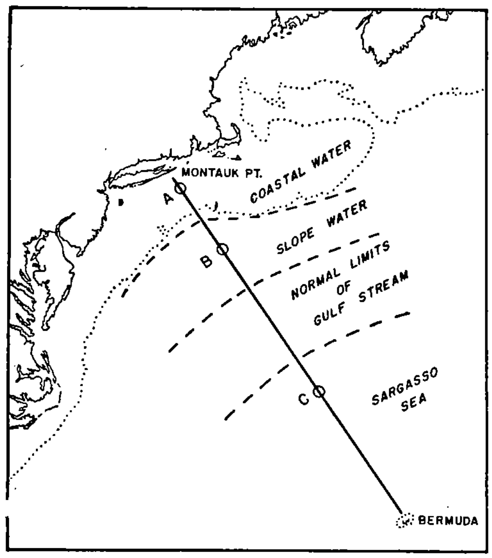

FIG. 1. Location of Gulf Stream section along which the sub-surface distribution of temperature and salinity has been observed on 15 occasions since June 1937.

On this line three points were selected at which especially complete and more frequent observations would be made. The first of these (station A) is about halfway across the continental shelf. This seemed representative of the coastal waters off southern New England and it appeared likely that the details of the seasonal changes in such a situation would prove interesting. The other two stations were located for the particular needs of the Gulf Stream investigation; station B at a point about halfway between the 100 fathom curve and the mean position of the northern edge of the current and station C some 250 miles nearer Bermuda. It was hoped that by making repeated lowerings on each occasion at these two stations, and thus partly averaging out the short-period changes, we would obtain good comparative values for the transport of the current. For this reason during the first six months of the program it was not considered essential to occupy stations at intermediate points each time that the *Atlantis* crossed the Gulf Stream.

To date 15 complete Gulf Stream sections have been occupied and on 4 additional cruises observations were made only at stations B and C. The dates and the station numbers of this primary part of the investigation are given on the next page.

This extensive series of subsurface observations is no doubt capable of yielding much information concerning many controversial problems in physical oceanography. However, the data will only be considered here as they bear on the variations in the Gulf Stream's flow.

The temperature and salinity profiles (Figs. 2–20) have all been constructed on the same scale, 1 nautical mile horizontally being equal to 5 meters in depth down to 2000

COMPLETE SECTIONS*

CRUISE No.	STATION No.	DATE
66	2859–2885	June 3–9, 1937
72	2905–2925	Oct. 3–10, 1937
74	3025–3037	April 6–13, 1938
77	3044–3069	May 29–June 8, 1938
77	3093–3112	July 9–17, 1938
80	3116–3147	Aug. 24–29, 1938
82	3155–3181	Sept. 29–Oct. 10, 1938
82	3200–3216	Oct. 27–Nov. 3, 1938
83	3217–3237	Jan. 5–11, 1939
83	3515–3537	May 28–June 3, 1939
84	3538–3559	July 6–12, 1939
87	3587–3611	Aug. 14–19, 1939
91	3684–3702	Oct. 5–9, 1939
92	3706–3724	Nov. 11–18, 1939
94	3747–3764	Jan. 18–23, 1940
	STATIONS B AND C*	
67	2894–2893	July 12–21, 1937
71	2902–2903	Sept. 13–18, 1937
73	2930–2931	Dec. 5–11, 1937
74	2943–2945	Jan. 3–7, 1938

* The deta ls of the observations at these stations are to be published soon in Vol. VIII No. 2 of this publication.

meters, and to 12.5 meters at greater depths. Thus the depth scale is exaggerated about 360 times in the upper part of the water-column.

While the complete series of temperature sections is presented here, for temperature most closely depicts the cross-current distribution of density,[3] only four of the salinity profiles have been included, selected to bring out some typical patterns found in the superficial layers. North of the Gulf Stream, below a depth of about 200 meters, and south of it at depths greater than 300 meters, the temperature-salinity correlation is extremely uniform. Thus except near the surface there is invariably very close agreement between the trends of the isotherms, isohalines and isopycnals on all Gulf Stream profiles crossing this sector of the current. However, some evidence recently accumulated indicates that in the western Sargasso Sea small but dynamically significant variations in salinity may occur at depths between 1200 and 1600 meters. It therefore has not been considered advisable in the transport calculations to determine densities on the basis of a fixed temperature-salinity correlation.

In working these sections an effort was made to complete the observations across the Gulf Stream in the shortest possible time, for we were less concerned with securing complete information at each station than in completing the observations between stations B and C as rapidly as possible. In a number of cases the weather was so unfavorable that the water-bottles failed to function properly on the deep lowering. On some of the sections this results in rather wide interpolations, which are indicated by breaks in the isotherms and isohalines. As will be explained later, these omitted observations from mid-depths have caused considerable difficulty in the dynamic analysis.

On the very first section (Fig. 2) made on the Montauk Point-Bermuda line (June 1937) an unexpected phenomenon was encountered. Well south of the Gulf Stream the main thermocline layer at two consecutive stations (Nos. 2873 and 2874) lay considerably nearer the surface than on either side. From a single section there is no possibility of

[3] We could, of course, have published density profiles. From the dynamic point of view this would have been a more logical choice. However, it is believed that the temperature sections are more interesting and for most readers have a more tangible meaning.

determining whether such a feature is caused by a huge internal wave, by two opposed currents, or by a powerful anticlockwise eddy. The navigational evidence, though not conclusive, gives some support to the eddy theory. Furthermore, general experience gained in studying several surveys of the waters just north of the Gulf Stream indicates that similar disturbances there are usually not greatly elongated in a direction parallel to the main current. Unfortunately no open network of stations has been occupied yet in the western Sargasso Sea. Therefore, it is with some hesitation that for the sake of simplicity such disturbances are spoken of as "eddies." The questions involved will be more fully treated in a subsequent report which is nearing completion.

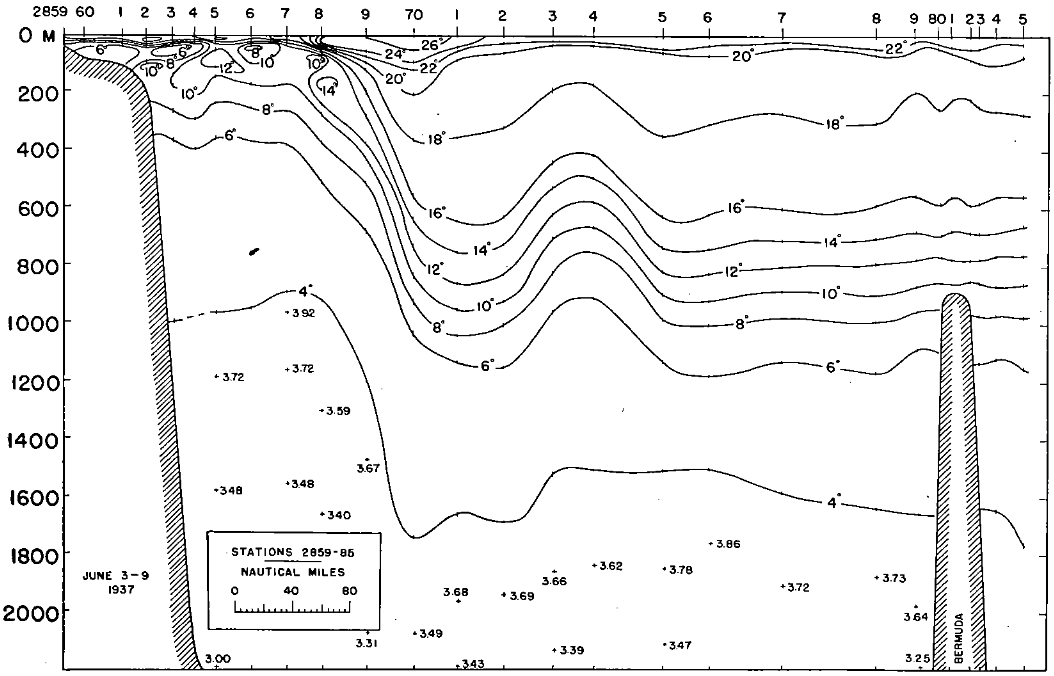

FIG. 2. Temperature section, Montauk Point to Bermuda, June 3–9, 1937.

At any rate it seems significant that if a straightedge is fitted to the trend of the isotherms at mid-depths south of this eddy, it closely coincides with the readings at station 2871. In other words, the eddy appears to have been superimposed on the gradual upward slope of the main thermocline layer south of the Gulf Stream, and hence that it did not influence the dynamic height on the right hand edge of the current. The much smaller unevennesses in the isotherms near Bermuda can be attributed to short-period internal waves, but the major disturbances, apparently can persist for periods of at least a month.

North of the Gulf Stream and near the surface this June profile clearly shows the complex mixing processes which are often encountered there. In examining this type of diagram it should be realized that the apparent lateral dimensions of such bodies of water of contrasting temperature are largely determined by the station interval.

The next complete section was not occupied until October, although during the summer stations B and C were visited on several occasions. This autumn temperature section

(Fig. 3) shows two prominent eddies, one north and one south of the Gulf Stream, which, under the simple interpretation used here, were apparently rotating in opposite directions. It will be noted that in both cases this rotation was contrary to the expected frictional effect of the Gulf Stream. Whatever the cause of these disturbances they have one serious consequence to this investigation. As long as the isotherms of the main thermocline layer on either side of the Gulf Stream can have such relatively large undulations, there is no possibility of determining transport values from a pair of fixed stations located at some distance from the mean path of the current.

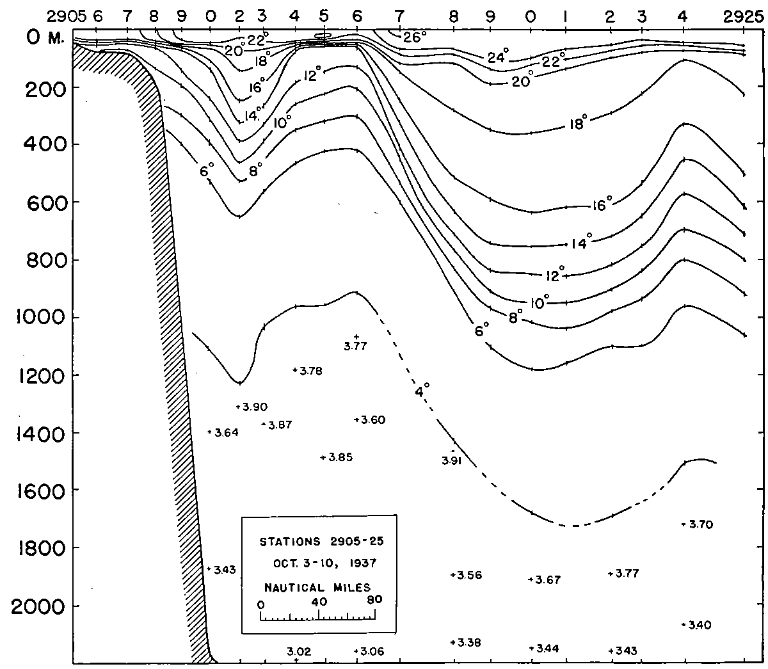

FIG. 3. Temperature section, Montauk Point to station C, Oct. 3–10, 1937.

The next complete section was not occupied until April 1938. Again the temperature diagram (Fig. 4) indicates two eddies, although the southern one is only partly shown. Due to a severe storm encountered as the *Atlantis* was crossing the Gulf Stream the station interval was too wide to locate accurately the northern edge of the current. It may be that station 3029 lay within the influence of the slope water eddy and, therefore, that its dynamic height was somewhat greater than would have been the case had observations been obtained 10 or 15 miles further south. For this reason the calculated transport from this line of stations in all probability is too low.

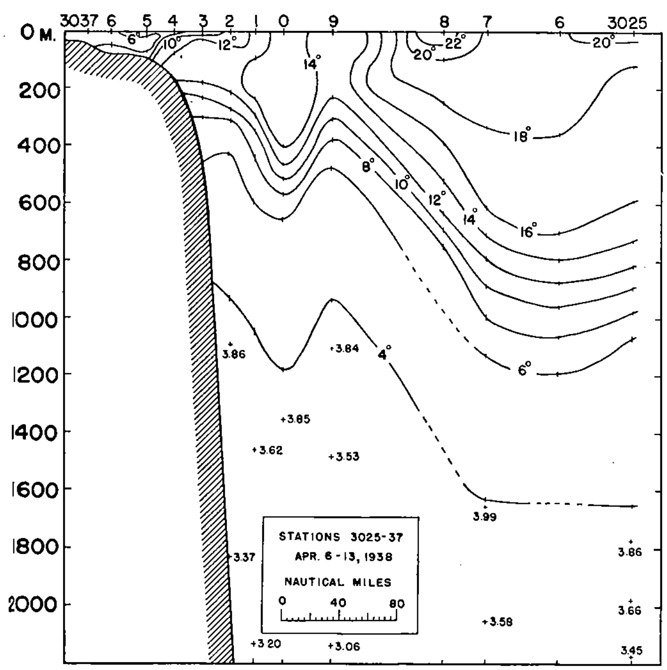

FIG. 4. Temperature section, station C to Montauk Point, Apr. 6–13, 1938.

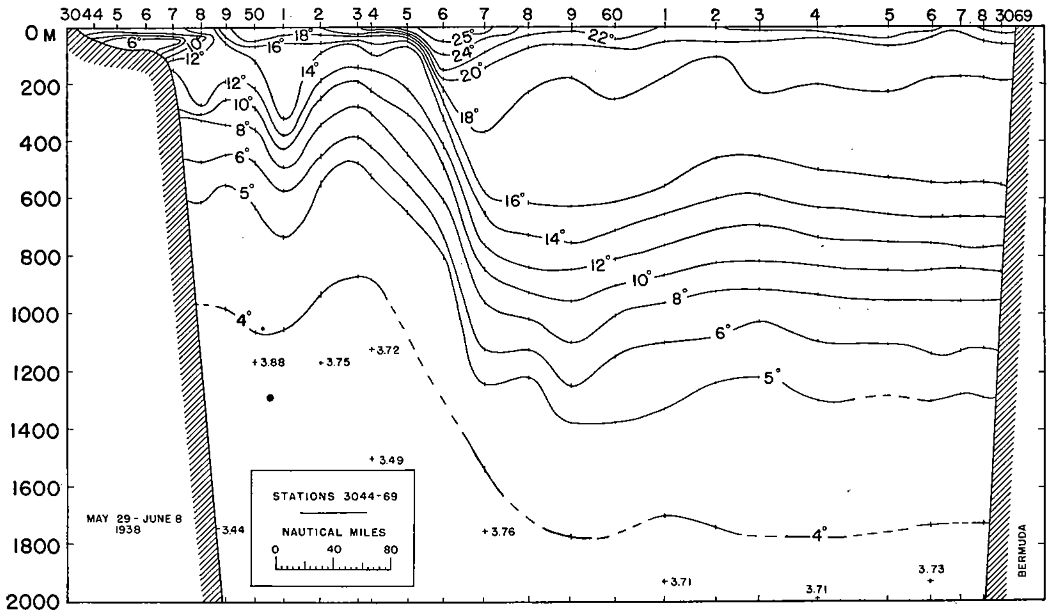

FIG. 5. Temperature section, Montauk Point to Bermuda, May 29–June 8, 1938.

The inadequacy of the observational program during the first year (June 1937–May 1938) is now only too obvious, for only three complete sections were obtained and on one of these the vital station near the northern edge of the current is missing. Moreover, on each occasion powerful eddies were present in the neighborhood of either station

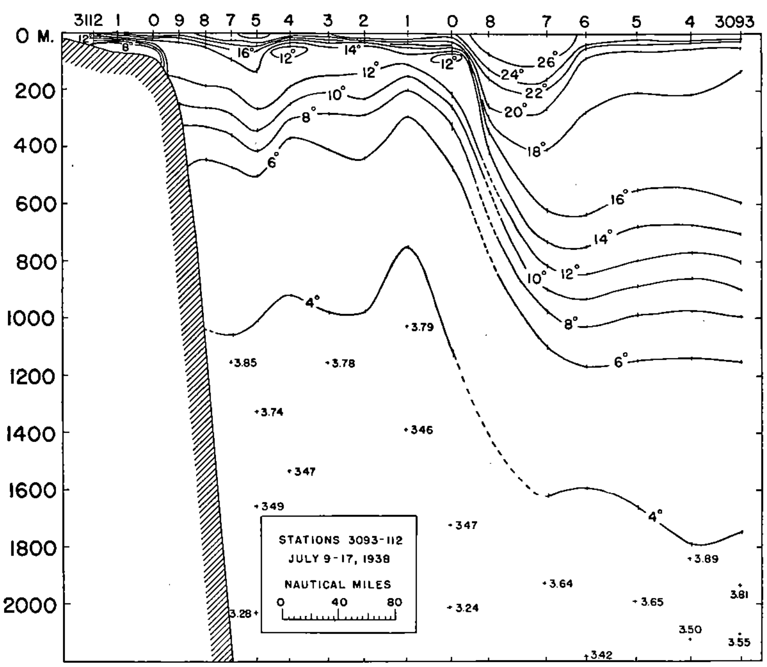

FIG. 6. Temperature section, station C to Montauk Point, July 9–17, 1938.

B or station C. Thus it is likely that on the four other occasions when only these two stations were visited, the observations were more or less affected by eddies and, therefore, fail to give the extreme cross-current density gradient. The results for the next year and a half are fortunately more convincing.

In early June 1938 the full section between Montauk Point and Bermuda was repeated. The temperature profile (Fig. 5) again indicates a well developed eddy at the axis of the slope water band, also a much broader but weaker disturbance south of the Gulf Stream, suggesting an eddy in its last stages.

A comparison between this section and the one made just a year before (Fig. 2) reveals that the axis of the current was about 30 miles further south in 1938 than in 1937. As will be shown later, this southward movement of the Gulf Stream was accompanied by an increase in transport. A month later (Fig. 6) the axis of the current had moved another 30 miles southward and the flow (page 24) had further increased. This rough correlation between the strength of the current and the distance of its axis from the American coast is discussed below (page 28) in connection with the thermograph records.

The eddy present in the slope water in June had nearly disappeared (or moved out of the section) by July (Fig. 6). On the other hand the weak Sargasso Sea eddy still persisted, though somewhat north of its former position. These large depth variations of the isotherms of the main thermocline layer on either side of the Gulf Stream can be interpreted in various ways, but our chief concern here is in emphasizing that, because from time to time such disturbances exist, a complete line of closely spaced stations must be made on each occasion in order to determine the extreme cross-current density gradient.

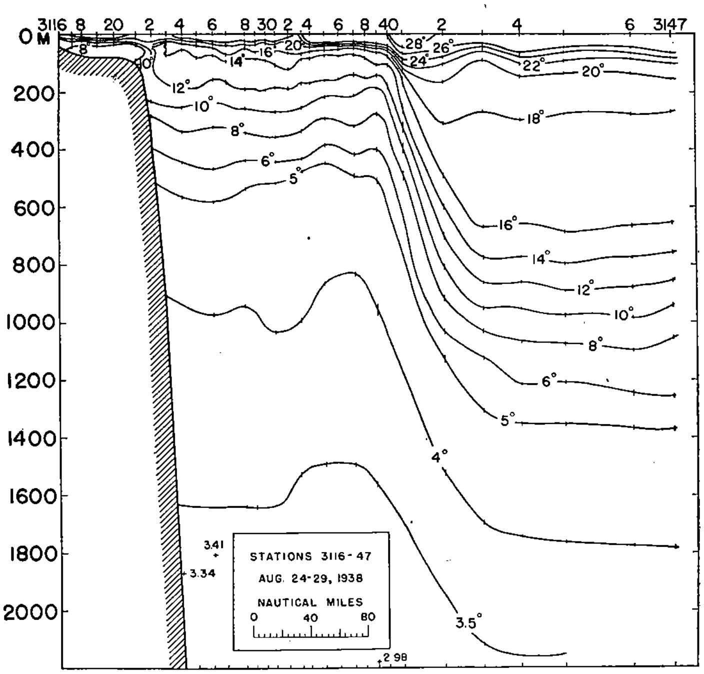

FIG. 7. Temperature section, Montauk Point to station C, Aug. 24–29, 1938.

The August 1938 section (Fig. 7) is in great contrast to the situations observed on the previous cruises, for relatively horizontal conditions were encountered both in the slope water band and in the Sargasso Sea. It will be noticed also that on the August section the station interval across the slope water band was exceptionally close. It is believed that the relatively small irregularities in the depths of the isotherms of the main thermocline layer north of the Gulf Stream indicate the magnitude of the short-period internal waves. Perhaps because of the close station spacing the northern edge of the Gulf Stream seems especially sharp on this diagram. However, towards the southern edge of the current the

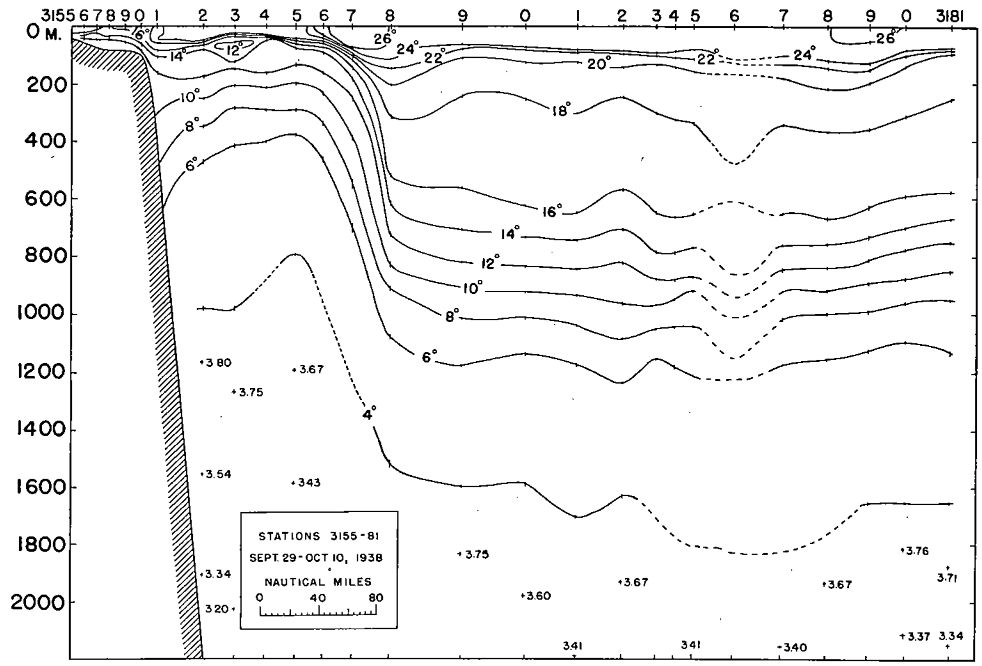

Fig. 8. Temperature section Montauk Point to Bermuda, Sept. 29–Oct. 10, 1938.

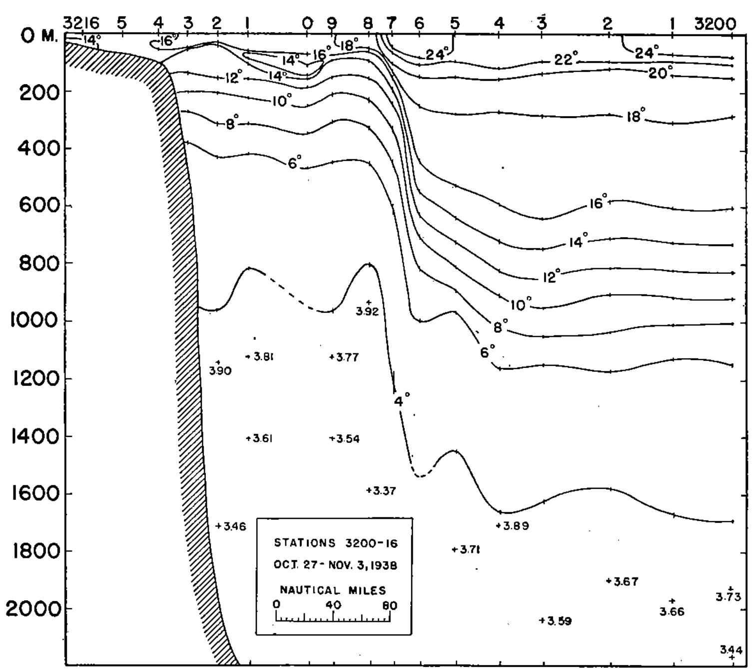

Fig. 9. Temperature section, station C to Montauk Point, Oct. 27-Nov. 3, 1938.

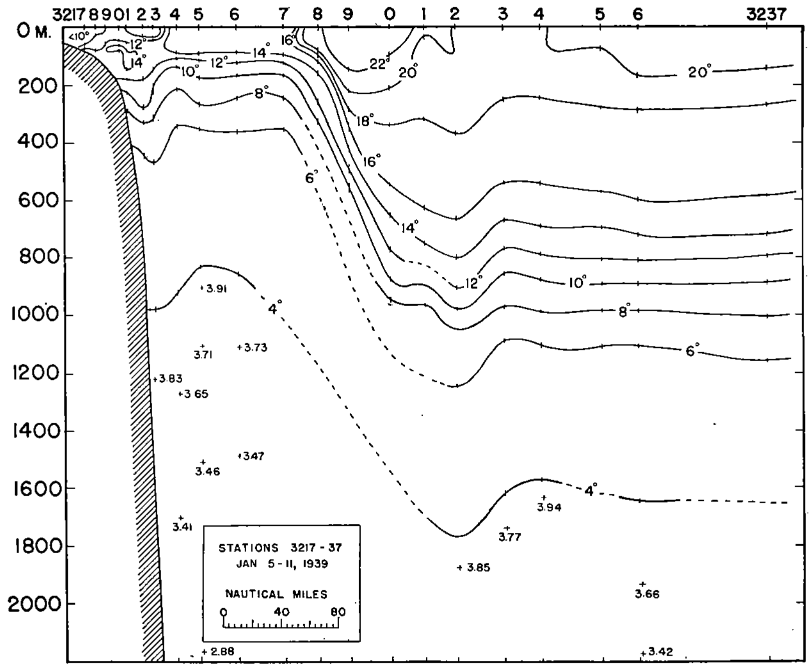

FIG. 10. Temperature section, Montauk Point to Bermuda, Jan. 5–11, 1939.

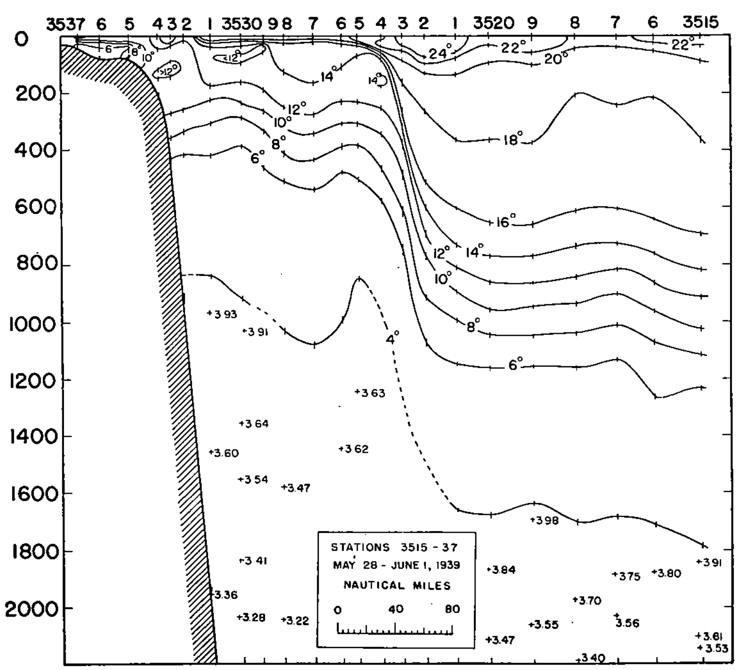

FIG. 11. Temperature section, station C to Montauk Point, May 28–June 3, 1939.

lateral gradient diminishes rather gradually. This indefiniteness of the right hand edge of the current introduces a difficulty in the transport calculations which cannot be solved in a satisfactory manner.

In early October, when the section was continued southward the whole way to Bermuda (Fig. 8) no marked eddies were encountered but the southern edge of the current remained very indefinite. The axis of flow had moved more than 40 miles north of its position in August, and even a superficial comparison of the two sections shows that the total cross-current thermal gradient, and hence the transport, was considerably less in October than in August.

On the return voyage from Bermuda, which began about two weeks later, the section was reoccupied from station C northward (Fig. 9). No change is evident in the state of the current and even the relatively cold pool of water near the surface at the axis of the slope water band seems to have persisted.

The first complete winter section (Fig. 10) was attempted in January 1939. Because of stormy weather and the swift current deep observations were omitted near the strongest part of the Gulf Stream. A more serious difficulty occurs in connection with station 3232 which was probably located near the southern edge of the current, for the five lower water-bottles seem not to have worked correctly. Likewise, the deeper readings at station 3231 are questionable and consequently the transport values obtained from this section are too low (page 27). During the winter months, north of Cape Hatteras, it is most difficult to secure an adequate set of observations from the Gulf Stream.

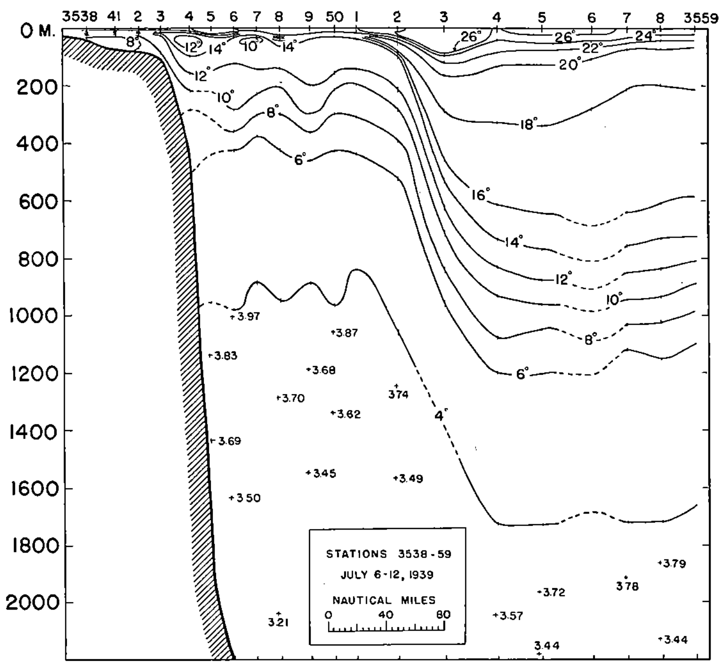

FIG. 12. Temperature section, Montauk Point to station C, July 6–12, 1939.

It was not feasible to occupy another section on the Montauk Point-Bermuda line until the end of May. This series of stations (Fig. 11), extending northward from station C, shows only moderate irregularities in the depth of the main thermocline layer on either side of the current. However, in this case, as during the previous summer (Fig. 7 and 8), the warm water lay deepest in the vicinity of station C, well south of the axis of swiftest flow. Except for the uncertainty which this introduces in the transport calculations, the observations were particularly successful.

Early in July a section out as far as station C was obtained. These observations (Fig. 12) reveal a completely normal arrangement of the isotherms. The warmest station was well north of station C and only slight unevennesses in the depth of the main thermocline layer were observed in the slope water.

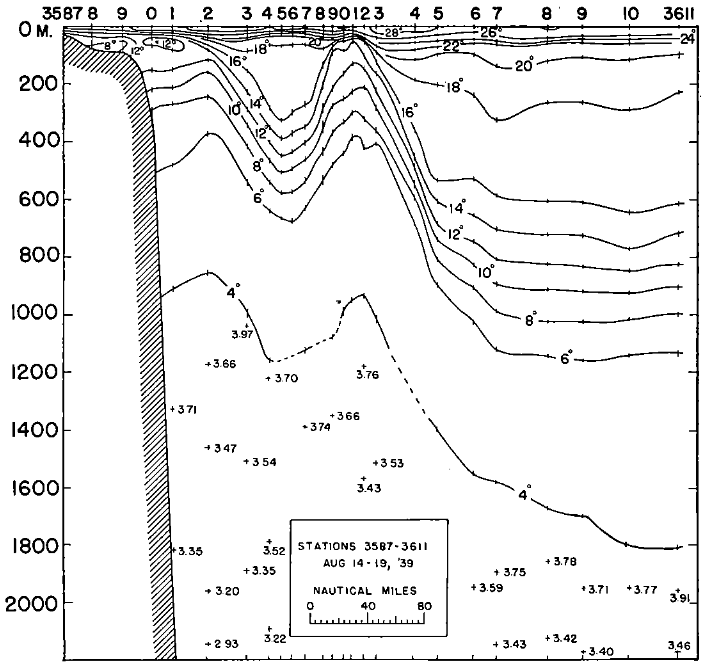

FIG. 13. Temperature section, Montauk Point to station C, Aug. 14–19, 1939. Above a depth of 100 m and between stations 3586 and 3591 the isotherms have been drawn for 4°C intervals only.

In August 1939 (Fig. 13) the stations north of the Gulf Stream crossed a well developed eddy, the first encountered in over a year. The lateral dimensions of this disturbance are particularly well substantiated, for six consecutive stations fell within it. The southern edge of the current was again rather indefinite, the isotherms at mid-depths continuing to deepen gradually out as far as station 3610. On this section a detailed investigation was made with a bathythermograph of the surface layer between the slope water eddy and the axis of the Gulf Stream. These results (Spilhaus, 1940) prove that even the 10 mile station interval from which a good part of Figure 13 is constructed is far too wide to disclose many significant details in the temperature distribution at depths shallower than 100 meters. However, in calculating the transport of the Gulf Stream it is the observations from deeper levels which are all-important and in this respect the *Atlantis* August 1939 profile was particularly satisfactory.

By October the large slope water eddy had disappeared, but the temperature section (Fig. 14) indicates that a smaller disturbance had developed close to the northern edge of the current, the band of swiftest flow being particularly narrow on this occasion. However, for the next 150 miles southward the isotherms at mid-depths continued to deepen gradually so that the whole width of the

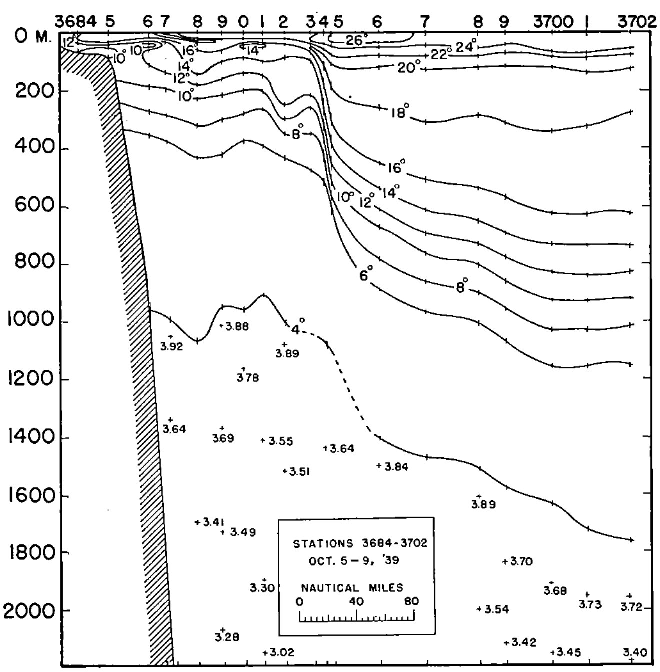

FIG. 14. Temperature section, Montauk Point to station C, Oct. 5–9, 1939.

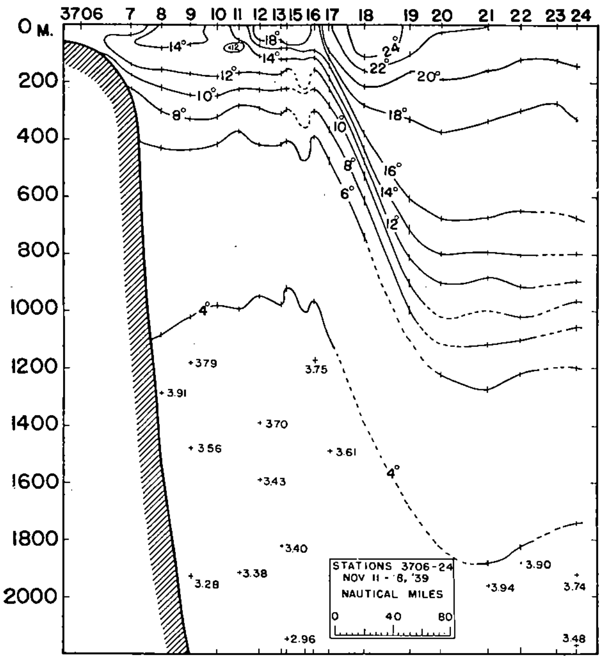

FIG. 15. Temperature section, Montauk Point to station C, Nov. 11–18, 1939.

Gulf Stream was much greater than normal. The observations at the two critical stations which mark the limits of the current (Nos. 3691 and 3701) were very complete and consequently the profile yields a reliable transport value.

On the November section (Fig. 15) at two stations the salinity observations were incomplete. However, it is clear that at these stations the temperature readings are reliable. Except for these mechanical difficulties the profile is satisfactory and shows a very normal arrangement of the water layers.

Finally, in January 1940, the most recent series of observations was obtained. Although the weather was no better than is usual in midwinter, the temperature profile (Fig. 16) is on the whole usable for our present purposes. At three stations (Nos. 3755, 3760 and 3761) there is some evidence that the deeper water-bottles may have tripped early. However, even when allowance is made for this mechanical failure, it is evident that the cross-current temperature gradient was very great. For example, the 10° isotherm descended to more than 1000 meters at the southern edge of the current, or nearly 40 meters deeper than on any of the previous sections.

The only features of salinity calling for emphasis are the following:

(1) As is shown by the June 1937 and the July 1938 profiles (Figs. 17 and 18) the distribution of salinity in the surface layer (down to 150 meters) of the slope water band is often considerably more complex than the temperature observations would indicate. This results from the mixing processes whereby coastal water of low salinity is being continually blended with water of Gulf Stream origin in this

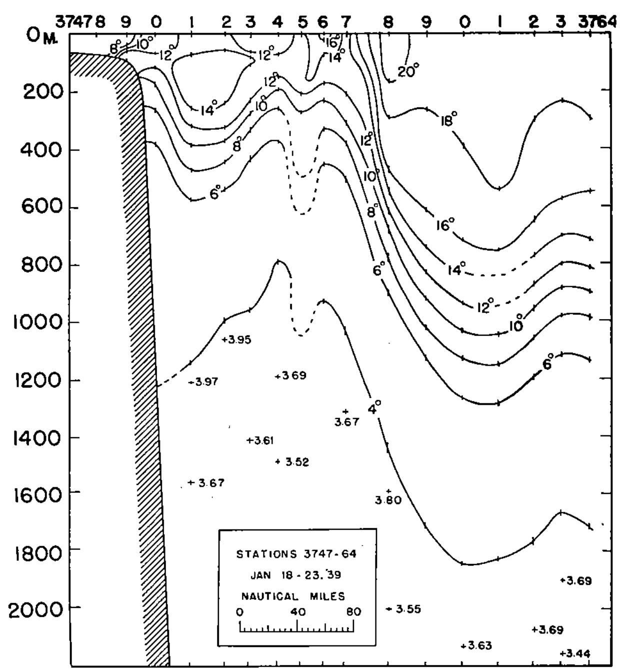

FIG. 16. Temperature section, Montauk Point to station C, Jan. 18–23, 1940.

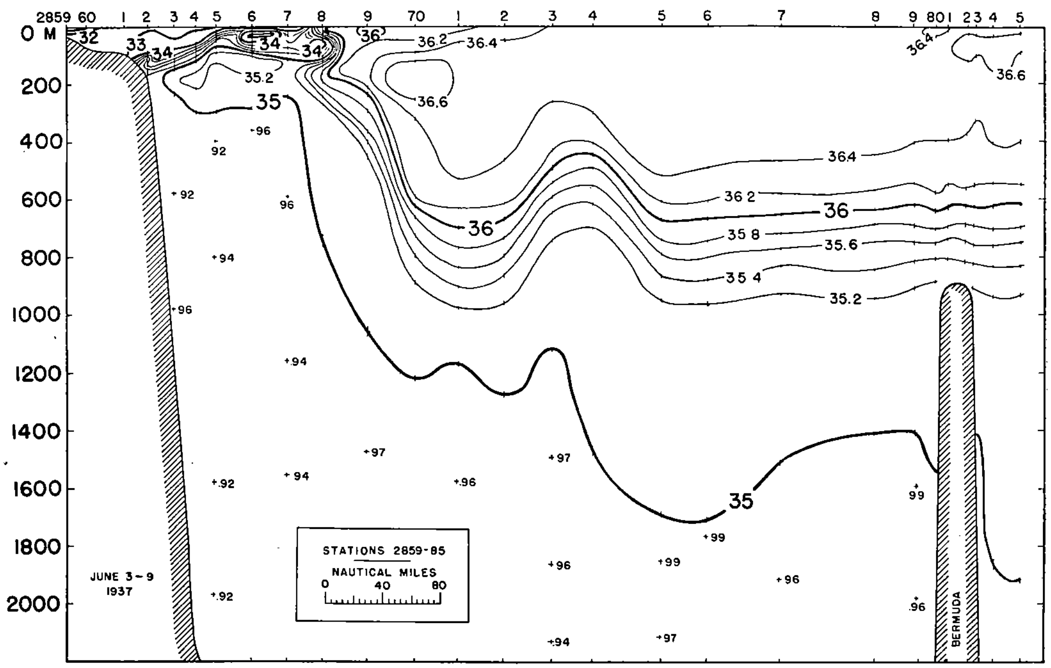

FIG. 17. Salinity section, Montauk Point to Bermuda, June 3–9, 1937. See Figure 2 for corresponding temperature observations.

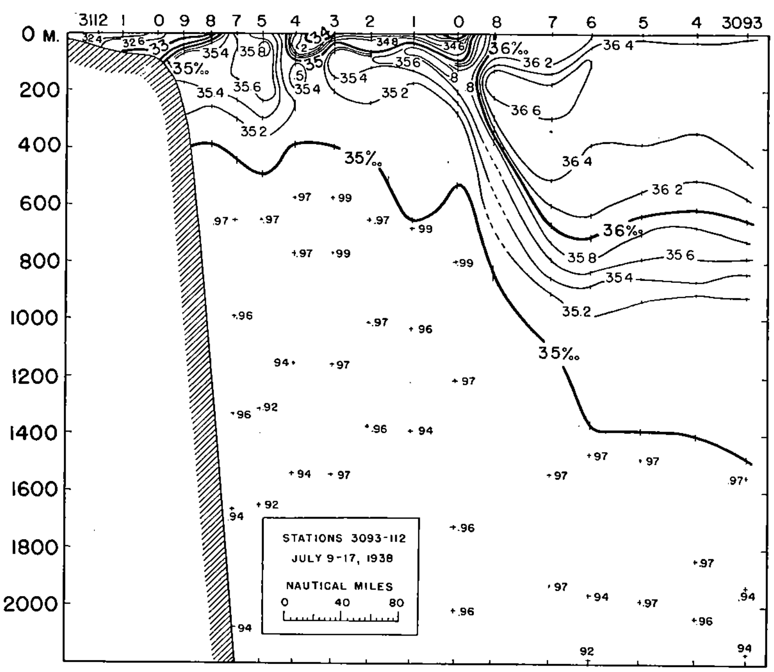

FIG. 18. Salinity section, station C to Montauk Point, July 9–17, 1938. See Figure 6 for corresponding temperature observations.

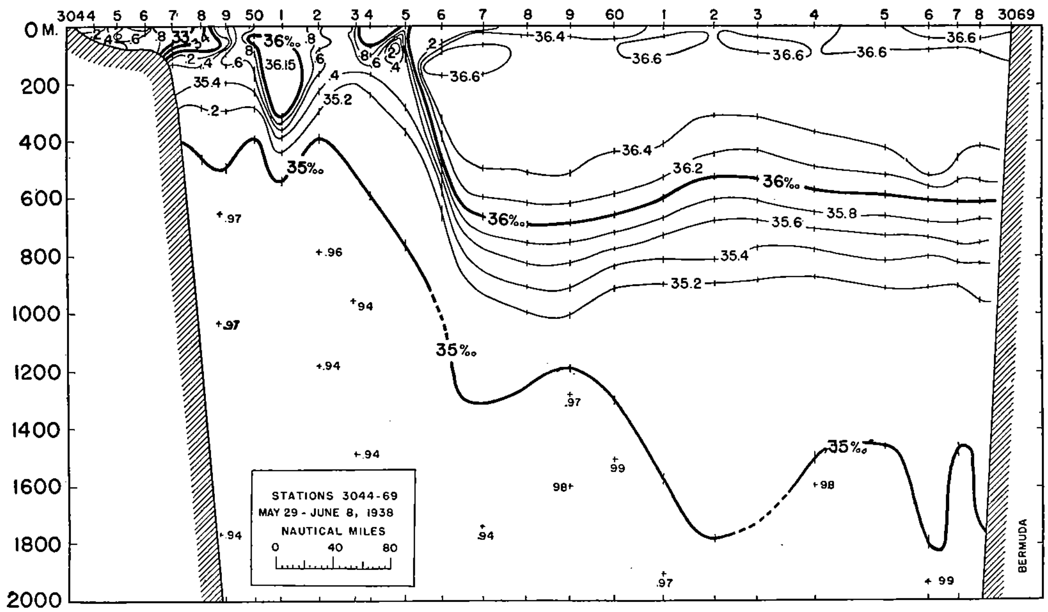

FIG. 19. Salinity section, Montauk Point to Bermuda, May 29-June 8, 1938. See Figure 5 for corresponding temperature observations.

region. However, the layer in which this rapidly changing temperature-salinity correlation is found is so shallow that the dynamic topography of the sea surface, relative to the 2000 decibar level, is only very slightly affected. Consequently it is not necessary to consider the superficial salinity pattern.

(2) Whenever a strong eddy was met with in the slope water, as was the case for example in June 1938 and August 1939 (Figs. 19 and 20) it surrounded a shallow core having salinities almost as high as those found in the Gulf Stream. In other words, these disturbances are associated with warm Gulf Stream water being carried well north of its usual position.

In the western Sargasso Sea, on the other hand, the eddies show no trace of Gulf Stream influence near the surface. True, the surface waters of the current, due to their southern origin, are always fresher than those of the western Sargasso

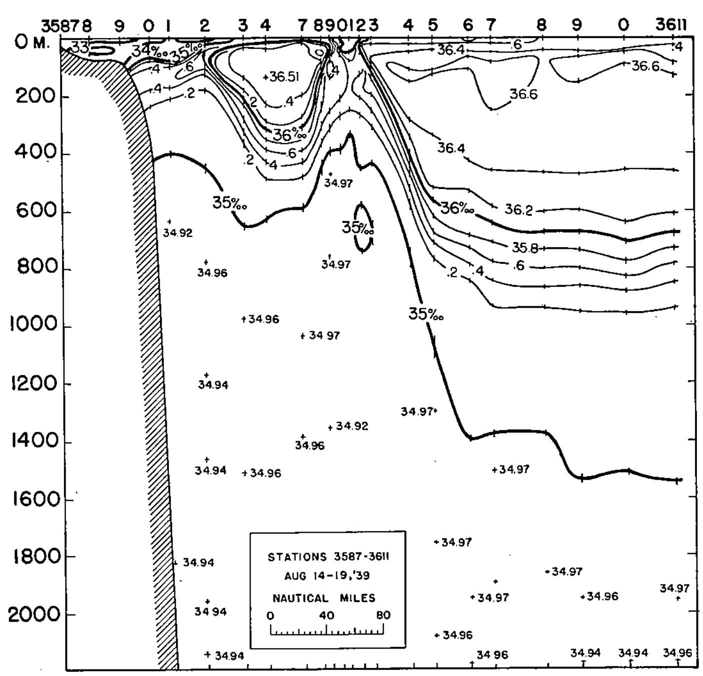

FIG. 20. Salinity section, Montauk Point to station C, Aug. 14–19, 1939. See Figure 13 for corresponding temperature observations.

Sea. But the difference is so slight that should Gulf Stream waters be drawn into an eddy beyond the southern margin of the current, it is doubtful that they would for long remain recognizable.

TEMPERATURE-DEPTH VARIATIONS AT THE NORTHERN AND SOUTHERN MARGINS OF THE GULF STREAM

To show up the seasonal cycle at the northern edge of the Gulf Stream and as a guide to the selection of observations for transport calculations, the slope water stations having the lowest temperatures at mid-depths[4] on each section are compared chronologically in Figure 21. This diagram also shows (by the broken curves) the results obtained on the occasions when only station B was occupied.

In selecting the coldest stations (the ones having the least dynamic height) from the 15 sections, preference was given in doubtful cases to those near the swift part of the current. When an eddy was present stations to the north of it were not used. In addition, some allowance has been made for the completeness of the observations. For example, station B, where the lowerings were repeated over a period of 12 hours or more, is obviously more representative than a nearby station at which only one lowering was made and which was perhaps only slightly colder at mid-depths because of the passage of a short-period internal wave.

It is evident that in selecting the station which marks the northern limit of the current on each section uncertainty is caused by a number of factors which are not easy to overcome with the present personnel and equipment of the *Atlantis*. If the temperature observations could be corrected and plotted within a few minutes after the water-bottles are brought to the surface, it might be possible for the scientist in charge to realize when he had reached the critical point on the section and to make certain that sufficient lowerings were obtained there to average out the short-period changes. However, because the northern edge of the current is usually somewhat inshore from the sharp rise in surface temperature, only when the section is run from south to north can the observer have an approximate idea of his position relative to the stream. The weather and the navigating conditions are only occasionally favorable enough to risk a twelve hour station at the very edge of the current, and when working within it each time the ship stops for longer than two hours she is more than likely to drift too far out of position. For these reasons it seems improbable that when using a single ship the water-bottle technique will ever be entirely satisfactory for locating the margins of a powerful current or for securing sufficient observations for the determination of the mean dynamic height at these two critical points on the profile.

However, in spite of these difficulties and restrictions the isotherms in Figure 21 are sufficiently regular so that the following tentative conclusions can be drawn concerning the seasonal variations in the waters at the northern edge of the Gulf Stream:

(1) The isotherms of the main thermocline layer (12° to 6°) have a tendency to be nearer the surface during the early summer (June, July and August) than at other seasons.

(2) A secondary peak appeared in January 1939, and less clearly in 1940. As will be shown later, this is corroborated by the coastal tide gauges and probably occurs annually.

(3) The main thermocline probably deepens in October or November and perhaps also in March or April. Thus it is possible that there are two maxima and two minima each year.

[4] In working with this material it has become quite evident that the depth of the main thermocline layer is a reliable indication of the dynamic height relative to the 2000 decibar level.

In addition, the diagram quite clearly indicates that for this sort of study observations should be made at least once a month. As has already been suggested, the April 1938 station places the isotherms too deep because of the eddy. Therefore, there are no reliable observations for the period January to May from either 1938 or 1939.

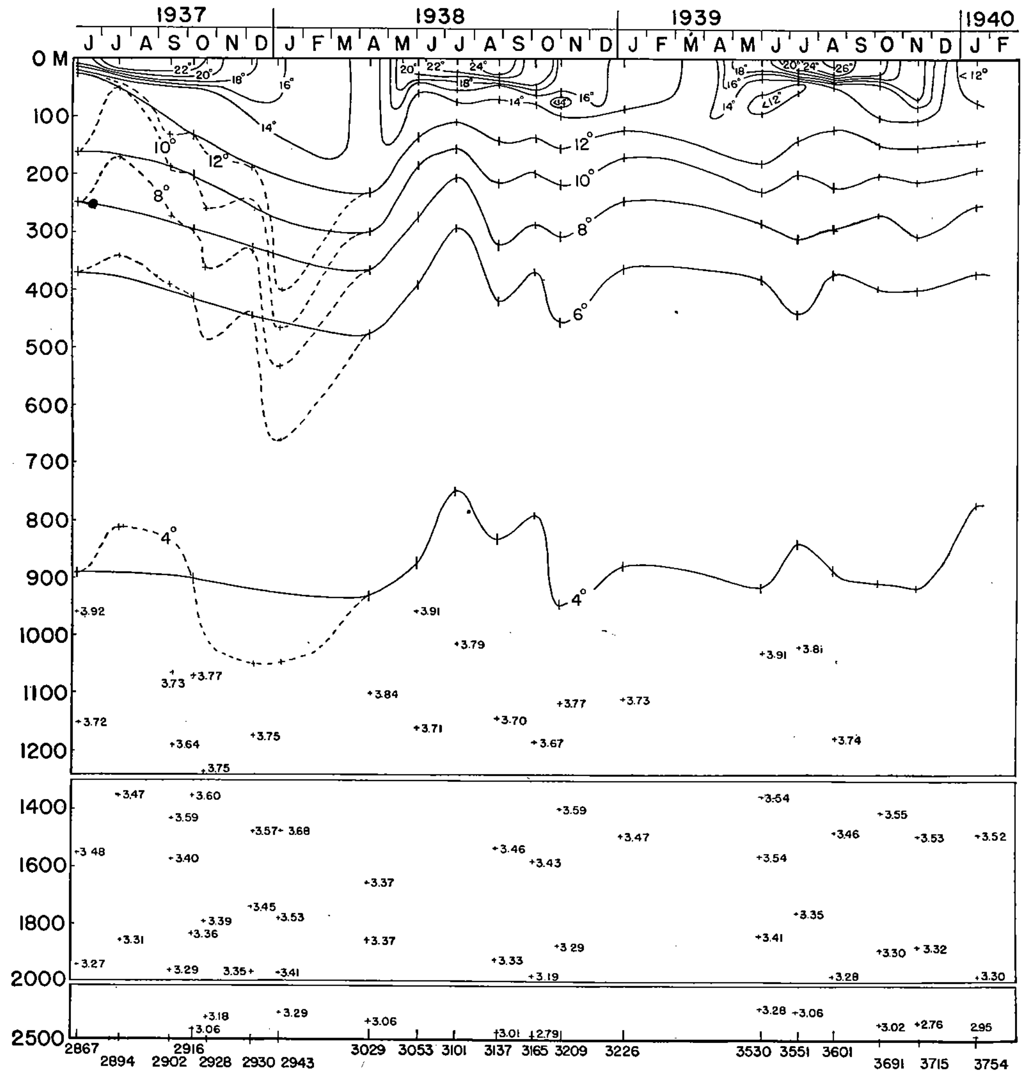

FIG. 21. Temperature-depth variations at the northern margin of the Gulf Stream. From each of the complete sections on the Montauk Point-Bermuda line the station at which the main thermocline lay nearest the surface has been selected and these temperature observations are plotted here against time.

The broken curves, which show the results when the data are included from the cruises on which only station B was visited, indicate a rather extreme maximum for July 1937, but the great deepening of the main thermocline in January 1938 was no doubt caused by an eddy. These observations have not been used for transport determinations.

The thermal changes at the stations which were nearest the southern edge of the Gulf Stream on each of the 15 complete sections are plotted in Figure 22. These are the stations where for the most part the main thermocline lay deepest. Because the velocity at times decreases very gradually in the direction of the southern limit of the current, the

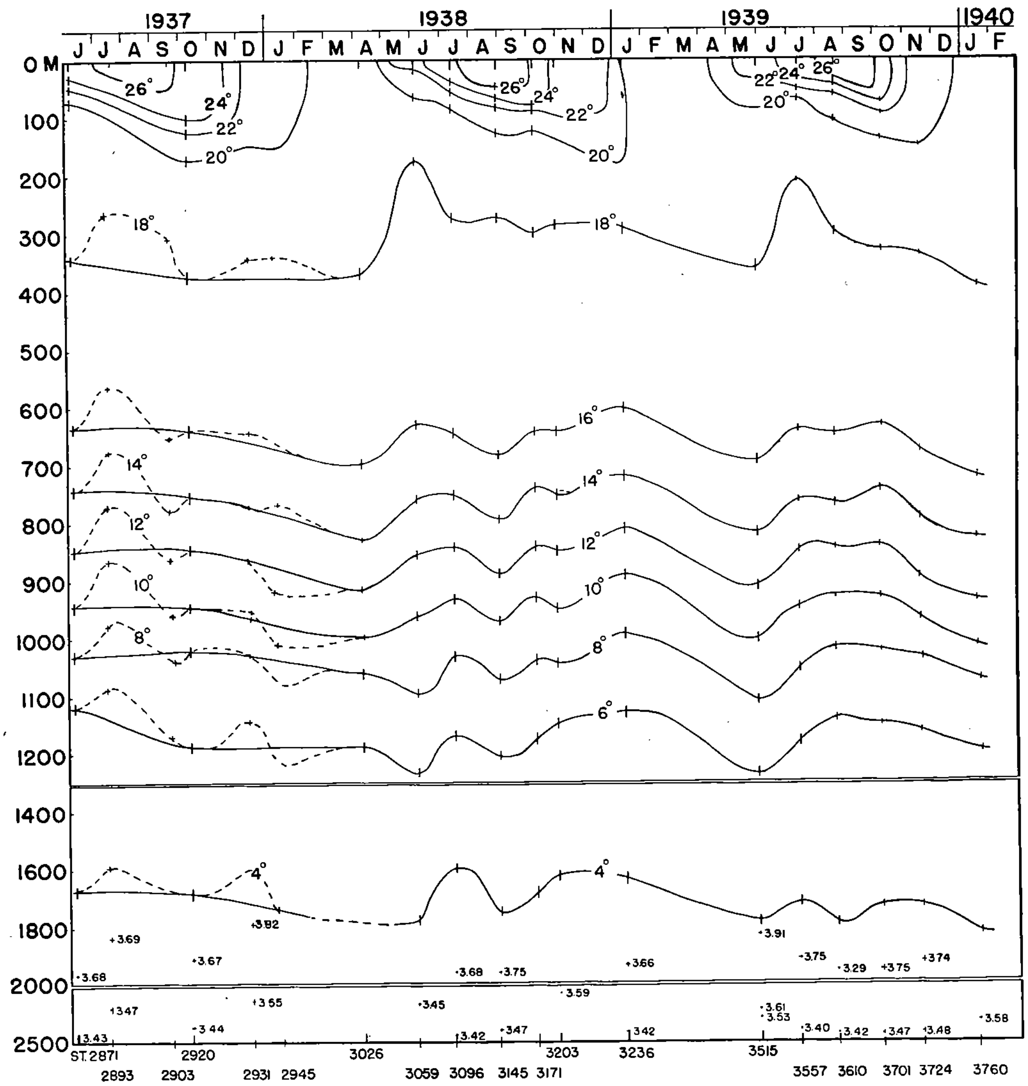

FIG. 22. Temperature-depth variations at the southern margin of the Gulf Stream. From each of the complete sections on the Montauk Point-Bermuda line the station at which the main thermocline lay deepest has been selected and these temperature observations are plotted here against time.

distance of these warmest stations from the axis of flow is even more variable than in the case of the coldest stations shown in the previous diagram.

The pairs of stations, one marking the extreme northern limit of the Gulf Stream and the other its southern boundary, which have been selected for the construction of

Figures 21 and 22 and for dynamic analysis are listed on page 24. The choice seemed obvious with only 6 of the 15 sections. On five of the sections the southern edge of the current was so indefinite that the completeness of the observations became the deciding factor. With two of the profiles (April 1938 and January 1939) the station interval near the probable boundaries of the current was far too wide, and on the remaining two sections it seemed necessary to use stations where some of the deep observations were missing.

If we disregard the broken curves, representing the changes on the occasions when only stations B and C were occupied, it is evident that the amplitude and phase of the temperature variations in the main thermocline layer at the southern edge of the Gulf Stream (Fig. 22) are somewhat the same as those observed at the northern edge (Fig. 21). There is a tendency for the main thermocline to be shallow in June or July, and deep in April or May and again in October or November. On both sides of the current the isotherms were shallow in January 1939, but in 1940 the southern station shows the opposite trend.

While it is obvious that the regularity of this seasonal cycle cannot be satisfactorily demonstrated from so few data, the diagrams at least indicate that, although in a good many cases the selection of the stations was rather arbitrary, the results are surprisingly consistent. In other words, the stations only once violate the general principle that when the main thermocline is shallow on the northern edge of the Gulf Stream it is also likely to be shallow at the southern edge and vice versa.

On theoretical grounds it can be argued that when the Gulf Stream decreases in transport the isopycnals at mid-depths, and therefore the isotherms, should deepen at the northern edge of the current and rise on the southern side. The fluctuations shown in Figures 21 and 22 are clearly not entirely explainable by this mechanism. It might also be claimed that, because of the smaller volume of water on the northern flank of the current off Montauk Point, most or all of the internal adjustment necessary for a change in transport should occur to the left of the axis of flow. While it is evident that the depth of the main thermocline has somewhat greater variations in the slope water than in the western Sargasso Sea, the density changes at mid-depths are, with only one exception (January 1940), in the same direction on both sides of the current. It seems possible that we have evidence here of long-period changes which may act over large areas of the ocean. The amplitude of these extremely slow oscillations seems to be about 100–150 meters. Superimposed on these gradual changes are the short-period waves which Seiwell (1937 and 1939) and others have studied. But in selecting the stations from the sections and by eliminating the effects of eddies the short-period internal waves do not obscure the more gradual changes.

The effects of changes in transport of the Gulf Stream are not at all evident in a superficial examination of Figures 21 and 22, for the cross-current thermal gradient at mid-depths appears to be relatively constant. However, the numerical analysis of these selected observations reveals significant fluctuations in transport, which moreover have a consistent trend.

TRANSPORT CALCULATIONS

Some evidence has already been given that the transport values which can be derived from the pairs of selected stations may not be as much influenced by short-period changes as might be expected on the basis of Seiwell's (1939) recent studies, for in the neighborhood of the two edges of the Gulf Stream the short-period internal waves seem to have a relatively small amplitude. In only a few cases was it possible to use either station B or station C for transport calculations. An examination of the repeated lowerings at these stations, where several series of observations were obtained, reveals that the short-period changes were indeed rather small.

The remaining selected stations were in most cases either the coldest or the warmest of the three or four available on each section in the neighborhood of the edges of the current. In using as far as possible the extreme stations within reasonable distances of the swift part of the Gulf Stream, the short-period changes have also been to some extent minimized. These transport values may all be too high, but at least they are roughly comparable from section to section.

In addition to the effect of short-period internal waves, the volume determinations are influenced, through the Coriolis factor, by the mean latitude used in the computations, because the pairs of selected stations varied more or less in position, depending on the width of the current and on its movements to and from the coast. As already pointed out, the positions of the southern stations are especially variable, being furthest north in June 1937 (Lat. 37° 08′) and furthest south in May 1939 (Lat. 35° 10′). Because of this fluctuating southern boundary of the current and the fact that the strongest flow was always much nearer its northern than its southern edge, it has not seemed consistent to use the mean latitude of each pair of selected stations in determining the transport. Rather, the latitude of the axis of strongest current (defined by the point at which the 10° isotherm reaches the 700 meter level) has been used. In doing this most of the transport values become somewhat smaller than would be the case if mean latitudes were used. The difference amounts to as much as 1.6% in the case of the January 1939 profile. However, it is believed that the results obtained are more comparable and more closely approximate the true volume of the current.

Besides the errors introduced by short-period internal waves and by the varying asymmetry of the velocity profile, the transport calculations are also influenced by small errors in the salinity determinations, especially at depths between 1200 and 2000 meters. As has been pointed out by Thompson (1939), this is particularly the case when the standardization reading is not correct, for then the mean density of the entire water-column may be considerably in error.[5]

The temperature-salinity correlation of the observations at the selected stations in waters having temperatures less than 4° is shown in Figure 23. The observations at the southern stations (crosses) fall, with few exceptions, within the band found typical of Gulf Stream sections off Chesapeake Bay (Iselin, 1936, Fig. 25), but at the northern stations (dots) the scattering amounts to as much as .07‰ on the salinity scale. Since it seems improbable that the observational error can be as large as this, it has not been considered wise to adopt a constant temperature-salinity correlation for the deeper layers in the transport calculations.[6]

This leads us to the base plane problem, for the effect on the calculations of slight variations in the salinity of the deeper layers could have been avoided by raising the base plane. The 2000 decibar surface has been used for two reasons. In the first place, as can be seen by examining the deeper observations plotted in Figures 21 and 22, the variations

[5] It is now clear that a considerable gain in the accuracy of the transport values could have been claimed had the water samples not been analyzed in the order in which they were collected.

[6] In the two cases where the selected stations did not extend deep enough, values for the lower part of the water column were interpolated from the stations on either side.

in depth of the main thermocline layer on either side of the Gulf Stream are paralleled by thermal fluctuations which extend downward to at least 2000 meters. Thus in all probability the flow at depths between 1500 and 2000 meters varies along with that of the superficial layers.[7] Secondly, if the base plane were set at a lesser depth than 2000 decibars, the calculated surface velocities would be less than are demanded by the navigational evidence, although they would show the same degree of variability as when the deep base plane is used.

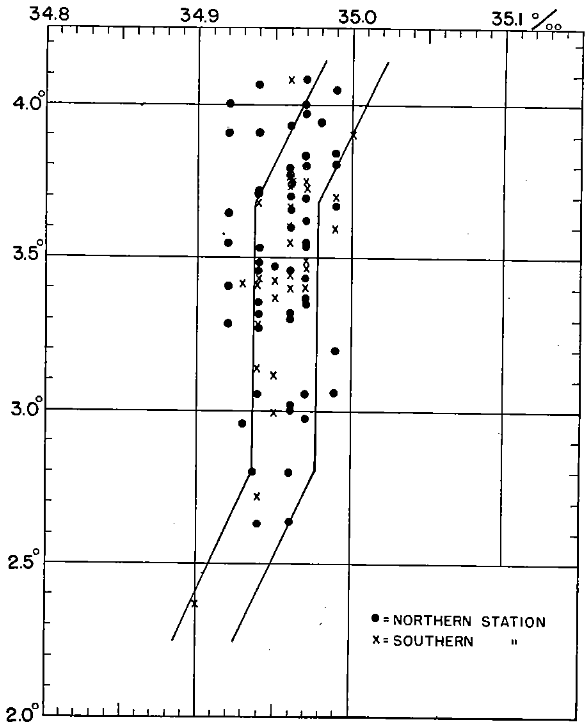

FIG. 23. Temperature-salinity correlation in the deeper layers at the stations selected for determining the transport of the Gulf Stream.

It must be admitted that, if all the various sources of error should happen to work in the same direction, the calculated sea level at such a station might be as much as 5 cm too high or too low. Thus when the transport is determined by a pair of stations, should both have a maximum error, but in opposite directions, the computed cross-current difference in sea level at the surface might be as much as 10 cm greater or less than it should have been. The mean surface dynamic gradient across this sector of the Gulf Stream (assuming the 2000 decibar level as motionless) is only about 110 cm, so it is conceivably possible that there could be an error of more than 10% in the calculated transport. But it is believed that the average error of the values[8] listed below is very much less than this.

STATION NO. COLDEST WARMEST	DATE	TRANSPORT IN MILLIONS OF CUBIC METERS PER SEC
2867–2871	June 5–6, 1937	83.1
2916–2920	Oct. 6–8, 1937	79.8
3029–3026	Apr. 7–11, 1938	78.7?
3053–3059	June 2–4, 1938	81.0
3101–3096	July 11–13, 1938	85.9
3137–3145	Aug. 26–28, 1938	87.4
3165–3171	Oct. 4–6, 1938	84.0
3209–3203	Oct. 29–Nov. 1, 1938	78.6
3226–3236	Jan. 7–10, 1939	78.2?
3530–3515	May 28–June 1, 1939	84.1
3551–3557	July 9–11, 1939	85.6
3601–3610	Aug. 17–19, 1939	80.8
3691–3701	Oct. 6–8, 1939	76.4
3715–3724	Nov. 15–18, 1939	80.4
3754–3760	Jan. 19–22, 1940	93.5

[7] While the computed velocity at these depths is small (less than 5 cm per sec) when the 2000 decibar level is used as a base plane, a shallower reference level reduces all the superficial velocities and thus has a marked effect on the calculated transport. For example, the transport referred to the 1000 decibar level is only about 67% of that found when the 2000 decibar level is used.

[8] In calculating the volume transport the values of σ_t at specified levels from the surface down to 200 m at the

It should be noted that in 1938 the August section gave the highest value, while in 1939 it was the July section. In both years the transport decreased to a minimum in October and none of the summer values is inconsistent with this sort of seasonal cycle. The doubtful determination from April 1938 has already been explained. In addition there is reason to believe (page 27) that the January 1939 value also is considerably too low. On the other hand, the January 1940 value seems suspiciously high. This may be partly due to a short-period disturbance in the vicinity of the northern selected station (see Fig. 16), but it should be pointed out that some of the occasional Gulf Stream sections occupied before 1937 yield even higher values.

If this investigation of fluctuations in the transport of the Gulf Stream had to depend entirely on the values obtained from the oceanographic sections, the results would hardly be impressive, for the reliability of the figures listed above are questionable from several points of view. At the outset of this study it was hoped that the oceanographic observations could be used to prove the usefulness of the tide gauge method. It now turns out, as will be discussed next, that the reverse argument can be better employed, for the tide gauge records indicate that 13 of the 15 transport values are surprisingly accurate.

TIDE GAUGE METHOD

It has long been recognized that changes in mean sea level, as observed by tide gauges, can be produced by variations in the strength of the nearby currents, but until recently there existed no oceanographic observations to indicate whether or not this was ever the predominating factor. The first such study was made by Montgomery (1938) who concluded that in the Straits of Florida and as far northward as Charleston, South Carolina, mean sea level along the coast is indeed rather closely connected with the strength of the Florida Current.

The pertinent tide gauge data will only be treated here in a preliminary manner, but in view of the recently published papers by Sverdrup (1938) and LaFond (1939) there now seems to be little doubt that in favorable locations a tide gauge can become an important instrument for dynamic oceanography.

As the Gulf Stream System flows northward from Miami towards Cape Hatteras, a strengthening of the current at the surface should in theory be accompanied by either a lowering of mean sea level along its western edge, by a rise of the sea surface on its eastern margin, or by both. In the Straits of Florida, where it is possible to set up tide gauges close to the boundaries of the current, the observations might be expected to yield a very satisfactory record of the variations in the flow. An instrument has been in operation since June 1931 at Miami Beach, Florida, and in February 1937, through cooperation of the U. S. Coast and Geodetic Survey, another was set up on Cat Cay, a small island almost directly across the channel.

For the period April 1938 to December 1939 the monthly departures from mean sea level at these two stations, expressed in feet, are as follows:

selected stations were scaled from temperature-depth and salinity-depth curves. Using Sverdrup's (1933) tables, the specific volume anomalies and the anomalies of dynamic height relative to the 2000 decibar level were next computed. Finally the volume transport in millions of cubic meters per second was determined by Jakhelln's (1936, page 4) method, with the exception that instead of using the mean latitude of each pair of stations the Coriolis factor was adjusted to the axis of strongest flow.

In view of the various sources of error discussed above and of the fact that we are mainly interested in relative changes in transport, the error pointed out by Ekman (1939, pages 27–28) in Jakhelln's calculation of Q, which amounts to 1%, has been neglected here.

Cat Cay

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1938	—	—	—	−.14	−.13	−.04	+.05	+.12	+.21	+.14	−.05	−.22
1939	−.27	−.22	−.07	−.01	−.03	+.10	+.18	+.22	+.33	+.22	−.20	−.19

Miami

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1938	—	—	—	−.22	−.09	−.14	−.26	−.05	+.16	+.54	+.60	+.10
1939	−.19	−.25	−.28	−.18	+.03	−.08	−.14	+.04	+.51	+.66	+.52	−.13

It must be taken into account that mean sea level at Cat Cay has been found as the result of only one complete year of observations, while the Miami station has been in existence for 8 years. As might be expected, these figures indicate that the changes in sea level are considerably greater on the left hand side of the current than on the right. At Cat Cay a change in mean sea level necessitates a readjustment of the currents to the eastward and, therefore, of sea level over the whole southwestern Sargasso Sea.[9] At Miami, on the other hand, only a relatively small volume of water is required to make the adjustment in mean sea level which must accompany gradual changes in the strength of the surface discharge through the straits.

Mean sea level is of course also influenced by the winds, by the barometric pressure and by the local seasonal variations in the density of the superficial layers. But in dealing with the tide gauge observations these additional variables will be to some extent neglected, for in various ways it is possible to eliminate them partly. For example, by emphasizing the average seasonal trend of sea level for a considerable number of years, the variations in the local weather become unimportant. Especially in the Straits of Florida, where the variability of the weather (except for occasional hurricanes) is slight, it seems permissible as a first approximation to interpret the monthly mean sea level values in terms of the strength of the current.

If it be granted that this simplification is allowable, the changes in monthly sea level difference between Cat Cay and Miami can be interpreted in terms of the average monthly surface velocity between these two stations. Assuming the average cross-current slope at the surface to be 55 cm, the surface velocities varied as follows in terms of percentage of the mean velocity for the period:

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1938	—	—	—	104	98	106	117	109	102	78	64	82
1939	96	102	112	109	97	110	118	110	90	76	60	97

In order to convert these figures to transport an important additional assumption is necessary, namely, that the surface velocity is indicative of the strength of the whole current. This is valid only if no prolonged stretching or shrinking occurs in the thickness of any particular layer. As yet there exist too few sections across the Straits of Florida to show whether or not this is indeed the case. But since all layers of the current take part in the variations in transport on the Montauk Point-Bermuda profile, it does not seem too bold a step to argue that the variations in sea level difference between Cat Cay and Miami are caused by changes in the transport of the current.

If the surface velocity changes listed above are thus considered comparable to transport variations, the changes in the Straits of Florida are more marked than those indicated by the dynamic analysis of the oceanographic sections between Montauk Point and Bermuda. However, they show a similar rhythm for there is a maximum in July and

[9] It is also important to emphasize that mean sea level at Cat Cay is the result both of the strength of the surface current through the Straits of Florida and of the Antilles Current. As will be explained later, these two currents probably do not have parallel variations.

a minimum in November, and a secondary maximum appears in March. Such a seasonal cycle was also found by Montgomery (1938) for the flow between Charleston and Bermuda.

The seasonal cycle of the discharge through the Straits of Florida is also well illustrated when the average seasonal change in sea level along the southeastern coast of the United States is examined. It must of course then be assumed that the greater part of the adjustment in sea level, which accompanies the variations of the current, occurs on its western side. As shown above, the relatively small range in mean sea level at Cat Cay supports this contention. On Figure 24, therefore, we have compared the average monthly departures from mean sea level at Miami (1931–1939) and Charleston (1921–1939), denoting changes in the transport of the Florida Current, and the variations in the computed strength of the Gulf Stream off Montauk Point (assuming $83 \times 10^6 m^3$ per sec to be the average flow). Considering that the tide gauge observations show the mean variations for a considerable number of years, while the transport figures are for specific dates during the last two and a half years, the lag indicated by this diagram has little significance. Both lines of evidence seem to point to an annual strengthening of the Gulf Stream in early June. The summer maximum is apparently reached in July or August and is followed by a rapid falling off in volume that persists until October or November. Unfortunately no transport calculations are yet available to confirm the prolonged winter maximum called for by the tide gauge data. As already explained, there is reason to doubt the reliability of the April 1938 and the January 1939 sections and in view of the agreement between the tide gauge record and the other 13 transport values, Figure 24 seems to provide additional evidence that this is indeed the case.[10]

The tide gauge records from Atlantic City (85 miles south of New York) were also examined, but this station and others north of Cape Hatteras do not show a close correlation with the changes in mean sea level in lower latitudes. Not only is the seasonal cycle of density in the waters over the continental shelf more marked and the current much further from the coast, but north of Cape Hatteras the Gulf Stream begins to curve eastward. A current flowing towards the east or west in the deep ocean can, by shifting its track (and hence altering its latitude), change in transport without causing mean sea level to vary in the relatively motionless waters on either side. As it moves northward the increased Coriolis force allows its volume to decrease without altering the slope of the sea surface and vice versa. Thus unless a current is forced by bottom topography to keep in a fixed track, the tide gauge method should only be applied in cases of flow having a marked north-south trend. While a closer study of the influence of local winds and climate on variations in mean sea level might afford means of correcting the tide gauge records from such a station as Atlantic City, there is no way to overcome the difficulty that north of Cape Hatteras the current is free to vary in latitude.

Our confidence in the reliability of the tide gauge observations at such stations as Key West, Miami and Charleston for this study of Gulf Stream fluctuations has been given further support by an examination of variations in mean yearly sea level. These values are plotted in Figure 25, again in such a way that points falling below the mean (showing a rise in sea level) indicate a weakening of the current. Adjustment has been made to allow for the difficulty that these three instruments do not cover exactly the same period of years. In other words, the Miami record (1932–39) has been fitted to the

[10] The very high value for January 1940 cannot be so easily dismissed as an error. As will be shown later (Fig. 26) it is well supported by the tide gauge record of the previous month.

mean for the same period at Charleston. It will be seen that for the years 1932 to 1939, when records are available from all three stations, there were three periods of increasing flow and one quite prolonged and marked (1935–1937) period of weakening currents. This record of the recent trend of the Gulf Stream is supported by the available oceanographic data and by Montgomery's (1938) analysis of the fluctuations in sea level, Bermuda minus Charleston. That is to say, the hydrographic sections on the Chesapeake Bay-Bermuda line indicate that 1932 was a year of decreasing transport (Iselin, 1936, page 34), while the Montauk Point-Bermuda sections show that, at least for the period July to November, 1939 was weaker than 1938 (see Fig. 24). However, the diagram also

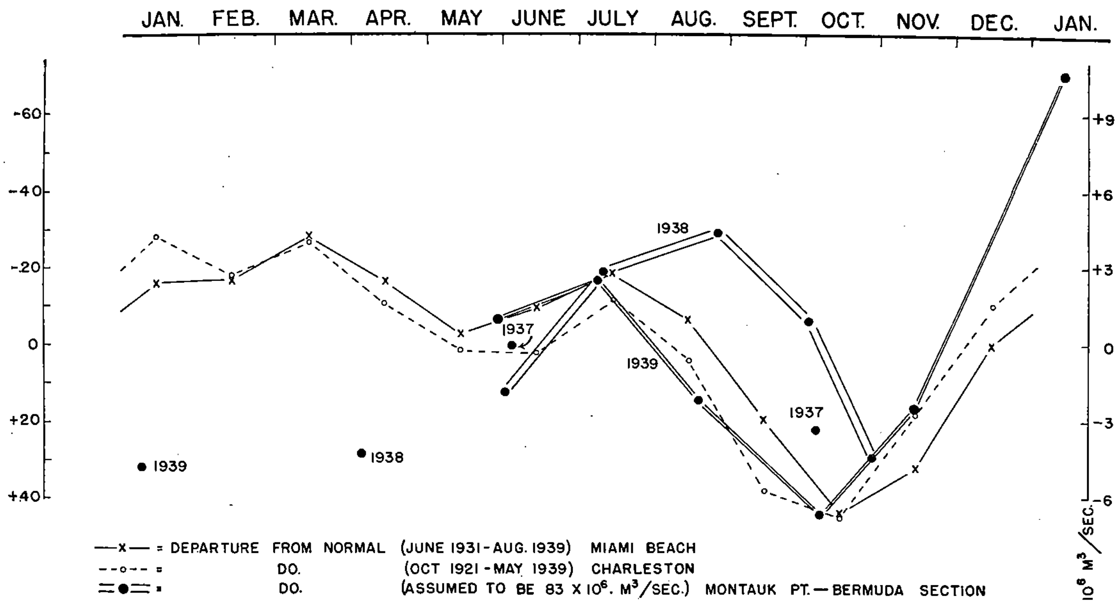

FIG. 24. Transport values derived from the oceanographic sections on the Montauk Point-Bermuda line compared with mean monthly sea level values at Miami and Charleston.

indicates that on the whole since 1932 the Gulf Stream has been relatively weak and that during the preceding six years it remained at or above the 1934 level.

If the tide gauge method is trustworthy, it appears that during the last 17 years the year to year changes in volume of the current have been slightly less pronounced than the average seasonal cycle. The latter produced a range of 0.6 feet in the mean monthly sea level values (Fig. 24), while the mean yearly figures (Fig. 25) varied by slightly less than 0.4 feet.

THERMOGRAPH METHOD

An analysis of the thermograms collected by ships of the Canadian National Steamship Company, running between Boston and Bermuda, has recently been published by Hachey (1939). He suggests on theoretical grounds that the Gulf Stream should weaken when its track moves northward and strengthen when the flow shifts southward. As discussed above, this can be expected also purely on the basis of the variation with latitude of the effect of the earth's rotation and thus the transport calculations would show this relationship if there were no changes in the cross-current density gradient. However, it is clear that the volume of the current fluctuates more widely than this.

From still another point of view variations in the transport of the Gulf Stream should bring about lateral shifts of its axis as it flows past New England. This theory (Iselin, 1938) rests on the assumption that the current forms part of a continuous, huge anticyclonic eddy, the central core of which contains a relatively fixed volume of warm water. Thus when the eddy increases in strength, causing a deepening of the main thermocline at the center of rotation, the diameter of the eddy should contract. A weakening circulation should then result in the Gulf Stream's pressing closer to the New England coast.

If this were the whole story, each thermogram from the Boston-Bermuda steamship track would directly provide a measure of the current's volume, for from these records, as Hachey has shown, the position of the northern limit of Gulf Stream surface water can easily be found. However, it seems likely that the main North Atlantic eddy migrates seasonally in a north-south direction following the wind system. Thus in winter, when the wind system is in a northerly position, the Gulf Stream will press close to the New England continental shelf in spite of the fact that at this season the transport of the current is at a maximum. From this standpoint the seasonal changes of the wind system are more or less opposed to the lateral shifts of the Gulf Stream which can be expected purely on the basis of the seasonal variations in its transport. This seems to explain why it is difficult to interpret Hachey's diagram (Hachey, 1939, Fig. 4) of the mean monthly northern limits of Gulf Stream water for the period 1933–1936 in terms of the annual cycle indicated by the transport calculations and the tide gauge records, discussed above.

If the thermograph records from the Boston-Bermuda track covered a longer period of years, and especially if such records were available from a more easterly section where the track of the current is probably more variable, it might be worth while trying to overcome this difficulty by assuming a regular cycle for the north-south shift of the center of the eddy and by correcting the positions derived from the thermograms accordingly. However, the curves given by Hachey (1939, Fig. 2) can be interpreted much more easily on a year by year basis. In this case the effect of the seasonal migration of the anticyclonic wind system over the North Atlantic can be neglected. Even a superficial examination of the curves shows striking differences from year to year. In 1934, for example, the position of the northern edge of Gulf Stream water was very constant and even during the summer warm surface water was not observed north of Lat. 38° 36′N. The opposite extreme is illustrated by the records from 1937 when the distance of the current from the edge of the continental shelf varied by 190 miles and when its mean position for the whole year was Lat. 38° 46′N. instead of Lat. 38° 03′N.

It seems likely not only that a weak current moves nearer the coast, but also that the northern limit of warm water becomes much more variable at the surface than when the flow is strong. It can also be argued that, during periods of weak flow, warm surface water is discharged from time to time to the left of the current. On the thermograph records this water is indistinguishable from the surface waters over the current.

Using both criteria, mean distance off shore and variability, it is easy to classify the various years, 1932–1938, for which Hachey has constructed curves giving the northern limit of Gulf Stream water on the Boston-Bermuda steamship track. Arranged in the order of decreasing flow this analysis of the thermograms gives the following results: 1934, 1932, 1938, 1933, 1936, 1935, 1937. Reference to Figure 25, demonstrating the departures from mean annual sea level at Charleston, Miami and Key West, will show that the same classification is called for by the tide gauge method.

The evidence provided by the thermograph records should not be given too much weight until a thorough analysis has been made of a longer series of data. Nevertheless, it gives some support to our belief that considerable confidence can be placed in Figure 25, even though the great majority of subsurface oceanographic observations now available are from years when presumably rather weak currents prevailed. In this connection the *Atlantis* sections of 1932 off Chesapeake Bay may also be significant. According to both the tide gauge record and the thermograms this was a year when the currents were about average in strength; in other words, considerably stronger than during the period

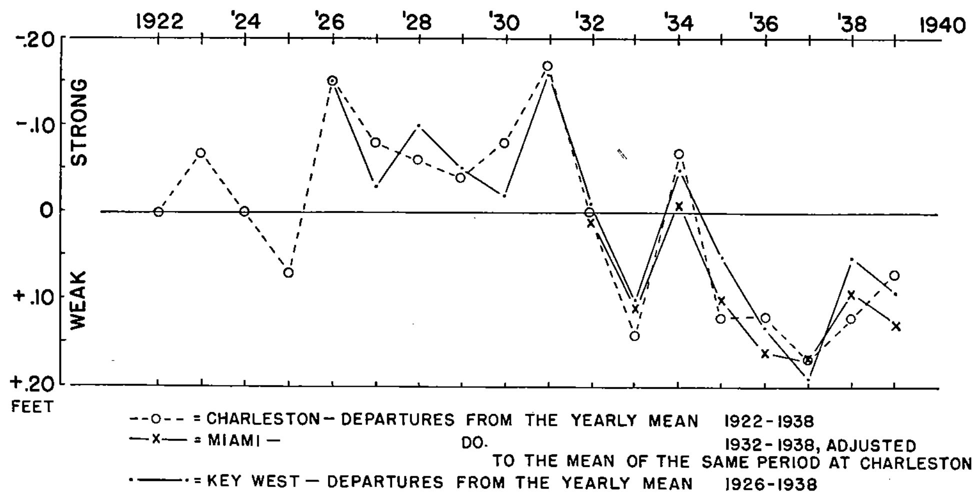

FIG. 25. Changes in mean yearly sea level at Charleston, Key West and Miami plotted in such a way as to demonstrate the year to year variations in the transport of the Gulf Stream.

covered by the oceanographic sections off Montauk Point. Especially, the February 1932 profile yields a transport value in excess of any obtained in recent years.

ROLE OF THE GULF STREAM

Evidence has been presented to show that the transport of the Gulf Stream, as it flows past New England, has varied seasonally and to a lesser extent from year to year during the short period June 1937 to January 1940. In addition, the yearly mean sea level values at Key West, Miami and Charleston since 1922 point to somewhat greater yearly changes over a longer period. An examination of the early Bermuda tide gauge records (1833–1843) confirms this view and suggests that over a sufficient length of time the long-period changes might approach or surpass the seasonal fluctuations in magnitude. However, this evidence is by no means conclusive and it is best to await additional oceanographic observations before speculating further on the extent of the long-period trends.

Before presenting a final interpretation of the transport calculations derived from the Montauk Point-Bermuda sections, our case will perhaps be strengthened by a discussion of the main warm water circulatory system of the North Atlantic. This picture of the role of the Gulf Stream has developed in the course of this investigation and is somewhat different from any previously suggested. It is no doubt a much simplified picture, but at present it seems to form a satisfactory basis for explaining the very fragmentary oceanographic observations as yet available for analysis.

Our primary simplification is the assumption that in the North Atlantic the main current system is comparable to a single enormous clockwise eddy within which the warm and relatively light water is found in the form of a huge lens (Iselin, 1938). According to this view, on the western side of the eddy the flow is more rapid and thus the central core of light water is not symmetrical, but has its greatest thickness at a point slightly west of Bermuda. In profile then the distribution of density on any northwest-southeast trending section may be schematically represented as in Figure 26 (situation A).

The total integrated effect of the anticyclonic wind system is furthermore assumed to be the major cause of this eddy and since the trade winds are so much steadier than the westerlies, it is considered that they provide the greater part of the energy. If the wind system remained steady in strength and position, there would be no direct cause for fluctuations in the transport of the continuous clockwise current surrounding the core of central, light water. However, the diameter of the eddy would not remain constant and the system would from time to time get out of adjustment unless the production and arrival of light water in the surface layer happened to balance exactly the outward lateral transfer effected by internal mixing.

It seems unlikely that turbulence can carry away the light water as rapidly as it is produced so that a more or less steady mass discharge takes place at the surface,[11] for the eddy cannot expand indefinitely in diameter. Under this view the excess of warm water is carried by the winds mainly towards the northeast, that is, towards the coast of northern Europe. This variable warm current, some of which finally enters the Norwegian Sea, is probably shallow and east of the Mid-Atlantic Ridge has a component somewhat to the left of the lines of flow of the main deep eddy. In short, even with a steady, fixed wind system the simplified eddy could not be expected to remain in balance without a more or less steady, northward, mass discharge of the central, warm surface waters.

However, that this discharge must have a large seasonal variation is at once demanded, if we take into account the fact that the winds are stronger in winter than in summer. Thus the transport of the main clockwise current must increase in winter, necessitating under the circulation theorem of Bjerknes and the wind current theory of Ekman a deepening of the less dense water at the center of the eddy. Adjustment is possible in one or both of two ways. The diameter of the eddy can decrease or the northeastern discharge can be interrupted. Since the seasonal change in the strength of the wind system is rather marked, both factors probably work together. This seems to explain what LeDanois (1934) has called "the annual transgression of Atlantic water."

At the same time that the winds are increasing in strength they also move southward, and thus for two reasons we should expect the Gulf Stream north of Cape Hatteras to be further off shore in winter than in summer. That this is not the case is explainable by the unique role which the Caribbean and the Florida Straits play in the circulation of the North Atlantic. The funnel-like channel which is formed by the West Indies imposes a physical obstruction to the eddy driven by the anticyclonic wind system. If the winds were fixed in position, the friction of this channel would merely be added to the internal friction against which the winds are working. But because of the annual north-south migration of the wind system, the frictional force centered in the Straits of Florida has a

[11] Through vertical turbulence some deep water is also continually being absorbed into the superficial layers and must be returned to high latitudes.

marked seasonal effect on the transport of the Gulf Stream System. As described above, in summer when the winds are north a smaller share of the North Atlantic eddy need pass through the Caribbean than in winter. Thus the frictional drag on the southwestern side of the eddy is lessened and the Gulf Stream actually increases in transport, in spite of the decreased strength of the whole wind system. However, in the southern part of the Florida Current the strong midwinter winds also increase the flow and this imposes a double rhythm to the more northern parts of the Gulf Stream System.

This point of view is to some extent supported by the tide gauge records from Key West, Miami and Charleston. At Key West average sea level is lowest in March (indicating the strongest current) and then rises gradually until November. This is to say that at the entrance to the Straits of Florida the flow appears to be strongest in winter and weakest in the autumn with no secondary maximum or minimum.[12] At Miami on the other hand, where the tide gauge measures the strength of the discharge through the straits plus the strength of northwestward flow outside the Bahamas, we again find the lowest sea level in March (Fig. 24), but a secondary minimum, resulting from the strengthening of the Antilles Current, is seen also in July. At Charleston this secondary minimum is slightly more marked, because in a down stream direction the contribution from the Florida Straits becomes relatively less and less important. Thus the winter transport maximum on the Montauk Point-Bermuda line, though more prolonged, is perhaps in some years less pronounced than the summer maximum. It seems possible that the *Atlantis* sections of 1938 indicate such a case.

In this discussion of the role of the Gulf Stream in the circulation of the North Atlantic there has gradually been introduced an apparent inconstancy. It has been suggested that changes in mean sea level along the southeastern coast of the United States provide most of the adjustment necessary for variations in the transport of the Florida Current, while at the same time it has also been argued that pulsations in the strength of the North Atlantic eddy as a whole require variations in the thickness of the warm superficial layers at its center. But if all the adjustment in sea level be along the left edge of the current, there is less reason to expect a marked variation in the discharge of warm water towards the northeast. The storage effect at the center of the eddy would then become less important. Before long it is hoped that some observational evidence can be obtained to help solve the problem created by these conflicting ideas.

Meanwhile, a more or less satisfactory explanation depends partly on the fact that from the Straits of Florida to Cape Hatteras the current flows over the continental slope. Along the left hand edge of this sector of the great anticyclonic eddy there is no deep water which is motionless. Only the shallow band of coastal water separates the current from the shore. Under these circumstances the main thermocline at the western margin of the Florida Current is free to occupy whatever depth is required by the strength of the current.

In contrast, from Cape Hatteras northward and east of the Grand Banks, any change in the depth of the main thermocline layer at the left hand edge of the eddy involves a readjustment in a relatively large mass of deep water. The three typical cross-sections of

[12] Montgomery's (1937) study of the changes in sea level between Key West and Miami gave just the opposite result. His evidence points to the current through the Straits of Florida being strongest in July with no pronounced winter maximum. This contradiction perhaps results from the fact that sea level at Miami is not uninfluenced by the Antilles Current. There is also the possibility that variations in the flow through the Old Bahama Channel may upset the belief that the flow passing Key West must closely equal that passing Miami. However this may be, our main argument is not seriously affected. The westward movement through the Caribbean and the resulting discharge through the Straits of Florida need not be expected to fluctuate in phase with the Antilles Current.

the North Atlantic eddy are illustrated diagrammatically in Figure 26. It is believed that the slope water is in free communication with the northern part of the ocean. Thus it is only necessary to consider the fundamental difference between situations B and C shown in this diagram. At Miami or Charleston a decrease in the strength of the eddy (B′)

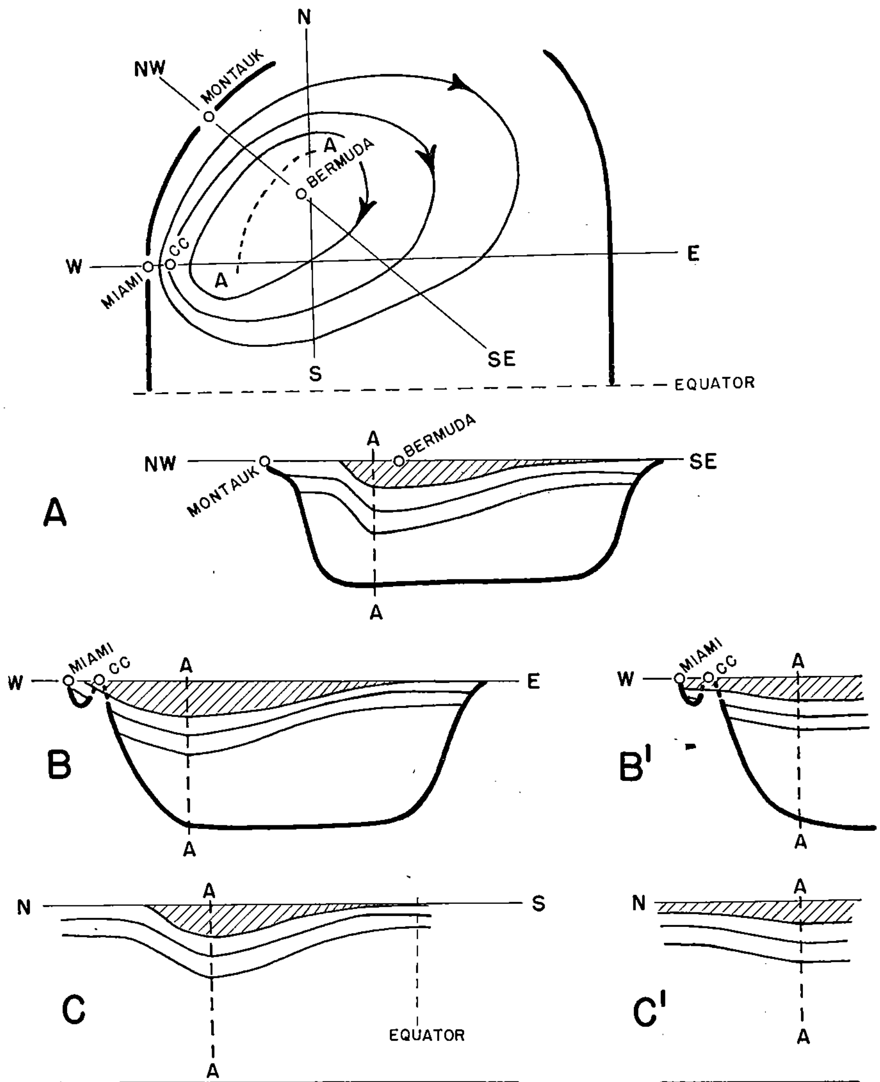

FIG. 26. Diagrammatic representation of typical density profiles crossing the main North Atlantic eddy at various points. For the sake of simplicity the sections show a three layered ocean, the lightest water being cross-hatched. Situations B′ and C′ show the change in density distribution which in theory must accompany a weakening of the Gulf Stream System.

requires only a slight cross-current shift of the warmest water, while off Montauk Point or in mid-ocean (C′) if the circulation weakens, a large area of the sea north of the current must be flooded with surface water of low density from the center of the main eddy. Thus, at least in the southwest, variations in transport need not much decrease the thickness of the warmest water. From this point of view it seems possible that tide gauges at Miami and Charleston are sensitive to quite minor variations in the current.

A possible explanation of the comparative stability of mean sea level at Cat Cay is also suggested by this diagram. Considering the eddy as a whole, Cat Cay is situated near the axis of the current. Thus sea level at Bermuda can be expected to show greater variations. Until a longer tidal record from both these stations has accumulated, this point cannot be satisfactorily tested, but it may be significant that for the period April 1938–December 1939 the extreme range of mean monthly sea level at Cat Cay was 0.6 feet while at Bermuda in the course of a year the range in the past has been as high as 1.03 feet.

The particular sensitiveness of the tide gauges along the southeastern coast of the United States is also evident from still another point of view. In mid-ocean should a minor change in transport occur, this can be taken care of by a lateral shift in the current, as explained previously, or by the development of a temporary counter current. It may be that a marked variation in the mass discharge of the warm central water occurs only as a last resort and as the result of quite large fluctuations in transport. Since off Miami and Charleston the current is flowing northward and since there is no room for the development of counter currents, sea level there must exactly follow the strength of the flow.

In all of this reasoning it has been assumed that the friction between the sea surface and the atmosphere is proportional to the wind velocity. However, there may exist critical wind velocities below which the winds exert relatively little force. Seasonal variations in air temperature and humidity, which in turn influence the stability of the lower layer of the atmosphere, may also play a part in the strength of the currents. For our present purposes the important point is that for several reasons we can expect a seasonal variation in the transport of the Gulf Stream. For part of the year the jet from the Straits of Florida is relatively strong and for part of the year the trade winds reach far enough north to force water freely westward between Bermuda and the Bahamas. What factors can be called on to produce fluctuations from year to year?

It is probable that the barometric gradient between the horse latitude belt and the equatorial belt does not vary greatly from year to year, in other words, that the average strength of the trade winds may change only very slightly from one year to the next. However, the effectiveness of these winds in maintaining the North Atlantic eddy may fluctuate quite widely. In certain winters, for example, the trades are much stronger in the western half of the ocean than in others. It may also be that sometimes the trade wind belt narrows and thus at its axis the wind velocities increase markedly. Especially if there is a critical wind velocity falling within the normal range of the trade winds, widely fluctuating currents may result.

Of recent years through a study of the pressure distribution meteorologists have been developing an index (zonal circulation intensity) which is a measure of the strength of the westerly winds of the northern hemisphere (Rossby, 1939). This circulation index is surprisingly variable and has considerable influence on the weather. With strong winds there is a more effective transfer of warm air to high latitudes and vice versa. Since the transport of the Gulf Stream is chiefly the integrated result of the torque imparted to the sea surface by the westerlies of high latitudes and the easterlies of low latitudes, it is suggested that the tide gauge observations at such points as Miami and Charleston provide an index of the general wind circulation over the North Atlantic.

It is hoped that this discussion of the causes and results of fluctuations in the strength of the Florida Current and the Gulf Stream and their role in the general circulation of the North Atlantic will have convinced the reader that the seasonal variations exhibited by the transport values computed from the *Atlantis* sections off Montauk Point are of the kind to be expected and that the tide gauge records from Miami and Charleston give a reliable picture of even quite minor variations in the Florida Current. It also seems reasonable that the changes in the Florida Current will be transmitted down stream, though how rapidly is as yet unknown.

In order to learn something about this lag and to fill in the very considerable gaps between the reliable determinations of transport from subsurface oceanographic ob-

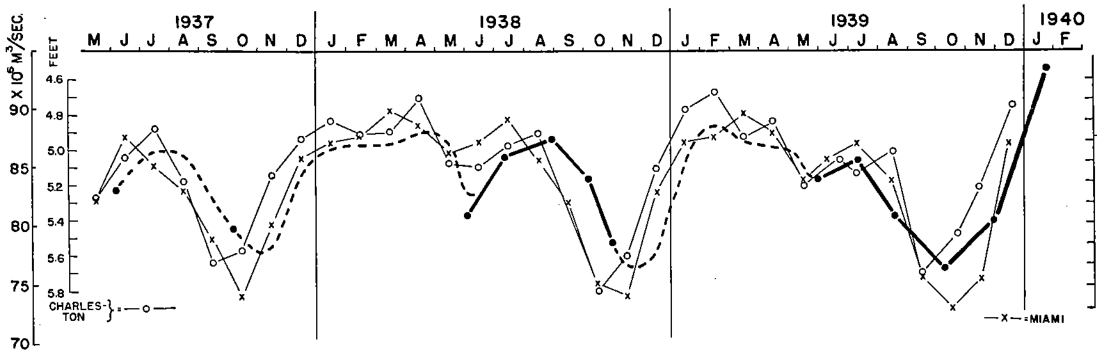

FIG. 27. The 13 reliable transport values from the Montauk Point-Bermuda section compared with the tide gauge record (monthly means) from Miami and Charleston for the same period.

servations, Figure 27 was constructed. The monthly values of mean sea level at Miami and Charleston were first plotted. When after several trials a suitable scale was found for the transport values from the Montauk Point section and the most trustworthy of them joined by straight lines, a lag of about two to five weeks seemed indicated. The broken curve, showing the strength of the current for the intervening periods was then drawn in with this lag in mind. It is believed that the diagram gives an accurate picture of the changes in transport of the Gulf Stream off Montauk Point since July 1937.

FUTURE COURSE OF THE INVESTIGATION

An obvious weakness in the final diagram is that to date insufficient Gulf Stream profiles have been occupied during the winter months. Two sections have been attempted in January, only one of which was successful, and none at all have been secured from either February or March. While winter observations are expensive, both in time and in equipment, it seems advisable during the next few years to fill in this gap so that the subsurface cross-current density gradient can be compared with sea level variations for all seasons of the year.

Meanwhile, the by no means weak circulatory system in the neighborhood of Bermuda is being examined by Dr. E. F. Thompson of the Bermuda Biological Station on the basis of the data now available. As soon as possible further observations will be secured from the *Culver*, the research vessel supplied to the Bermuda Biological Station by the Royal Society of London. This work is not only of interest in itself, but provides a most necessary background for the interpretation of the Bermuda tide gauge record.

It also emphasizes the weakness of one of our most vulnerable assumptions, namely that the Gulf Stream System and the Northern Equatorial Current form a continuous huge eddy with a large, relatively motionless central core. In an earlier report (Iselin, 1936) it was suggested that the Gulf Stream branches near the longitude of the Grand Banks of Newfoundland and that some of its waters return rather directly past Bermuda towards the area northeast of the Bahamas. For part of the year (summer) Dr. Thompson's investigation indicates that this is the case, but during the opposite season the predominating flow past Bermuda seems to be towards the northeast. If this change is only a local one, it will not greatly affect the theory developed in these pages. However, it is clearly desirable to extend greatly the range of the surveys which have been made in the central part of the Sargasso Sea.

Now, if the Gulf Stream, in addition to the part which escapes to the north, divides into several branches, one returning directly by way of Bermuda, another flowing past the Azores and a third towards the Norwegian Sea (Iselin, 1936, Fig. 48), then we may not be dealing with a simplified huge eddy at all. In such a system the fluctuations imparted to the Gulf Stream by the variable strength of the winds might be compensated for by changes in the relative strength of the two or more southerly branches. To settle any such question the whole area of the Gulf Stream System must be repeatedly surveyed, an undertaking clearly outside the resources at our disposal. However, the correctness of the theory can perhaps be tested more indirectly.

In the case of the waters over the American continental shelf between Cape Hatteras and the Grand Banks the theoretical influence of fluctuations in the transport of the Gulf Stream seems to be borne out on preliminary examination. For example, during years of supposedly extremely weak Gulf Stream flow, such as 1937 and 1939, when the current not only should move nearer the American coast but when it is also likely that warm water will be discharged to the north, surface temperatures near the coast and summer air temperatures on shore (because of prevailing southerly winds) were notably high. On the other hand, a year like 1934, when the tidal record requires a relatively strong current, was characterized by low inshore surface temperatures and a relatively cool coastal climate. These and other similar correlations between Gulf Stream transport and the distribution of temperature and salinity along the northeastern American coast have already received some study with encouraging results. The next step will be to apply the same reasoning in the case of European coastal waters. If the storage effect at the center of the North Atlantic eddy is as important a factor as has been suggested above, it should be possible to support the theory through a study of oceanographic and meterological records from the northeastern part of the ocean.

The objection will no doubt be raised that of recent years detailed oceanographic studies have more and more emphasized the complexity of the current pattern. However, such smaller scale phenomena seem not to be incompatible with the ideas developed in these pages. The basic movements of the warm and more powerful currents in mid-latitudes appear to be circular, though in detail they may be most complex. From this huge eddy a number of larger and smaller branches originate, the chief one being the most northern branch, the North Atlantic Current. Some of this relatively warm water, after it has crossed the Mid-Atlantic Ridge, flows northeastward to supply the Irminger Current and the Norwegian Atlantic Current. The strength of these currents and the temperature of their superficial layers are here held to be dependent on the strength of the main eddy in lower latitudes. It is argued that the supply of warm water approaching the coast of northern Europe will be increased by a weakening of the general wind system and that it may be greatly decreased by a sufficient strengthening of the Gulf Stream System. However, once the main eddy has become steady again, whether it be strong or weak, the normal transfer of warm surface water from mid-latitudes to high latitudes should be resumed at about the ordinary rate. In addition, of course, the surface temperatures off the European coast depend on meteorological factors, but it is hoped that on close examination it will be found that the oceanographic mechanism is not entirely obscured by variations in the local weather.

SUMMARY AND CONCLUSIONS

(1) On fifteen occasions since June 1937 a line of stations, extending across the Gulf Stream off Montauk Point, has been occupied by the *Atlantis*. These temperature and salinity observations form part of a cooperative investigation of long-period variations in the transport of the current and were undertaken at the suggestion of the Bermuda Oceanographic Committee of the Royal Society of London.

(2) If the fluctuations in transport of the Gulf Stream are to be computed from subsurface temperature and salinity observations by use of the circulation theorem, only complete sections of closely spaced stations can be expected to yield comparable results. It is unlikely that any considerable simplification of such observations can be developed to produce reliable values.

(3) For the period June 1937 to January 1940 the available subsurface observations indicate that the transport of the Gulf Stream has varied between a maximum of about 93 and a minimum of about 76 million cubic meters per second, assuming the 2000 decibar level as motionless. The sections analyzed dynamically indicate that the current is relatively strong during the early summer, falls off rapidly in strength to a minimum in October or November and then increases rapidly until January or later. A secondary minimum is called for in April or May, but it is not clear from the available transport determinations whether or not the more prolonged winter maximum usually surpasses the summer maximum in strength.

(4) These results have been compared with tide gauge records from Miami, Florida, and Charleston, South Carolina. On the assumption that the sea level difference across the current varies with the transport of the Florida Current, the monthly sea level values at these two stations not only corroborate the seasonal cycle indicated by the dynamic analysis but also show that in 13 out of 15 cases these results are comparatively accurate.

(5) If the mean yearly sea level values along the southeastern American coast reflect fluctuations from year to year in the strength of the Florida Current and hence of the Gulf Stream, the years since 1934 have been on the whole characterized by weak currents, while during the previous 10 years, notably in 1926 and 1931, they were relatively strong. This evidence also suggests that during the past 17 years the long-period fluctuations in transport have not quite equalled the seasonal cycle in magnitude.

(6) A study of the thermograph records collected by steamers running regularly between Boston and Bermuda during the years 1930–1938 gives support to the results derived from the available tide gauge records and from the dynamic analysis. The method used rests on the assumption that during periods when the Gulf Stream is weak, warm water will from time to time be carried north of its usual limit and vice versa.

(7) A tentative explanation of the agreement between these three methods, and of the role of the Gulf Stream in the circulation of the North Atlantic is presented. This ex-

planation depends primarily on the assumption that the major, permanent, warm currents of the North Atlantic are comparable to a huge eddy, from which various shallower offshoots, such as the Norwegian Atlantic Current, are derived. The Gulf Stream System forms the western and northern quadrants of the circuit, while the Canaries Current and the Northern Equatorial Current constitute the eastern and southern. Because of the seasonal variation in the strength of the anticyclonic wind system over this area and its annual north-south migration, the main circular currents will vary in transport. In winter, when the torque of the winds increases, the Gulf Stream System can be expected to strengthen. But in summer, when the winds move northward, a larger percentage of the Northern Equatorial Current is able to supply the Gulf Stream without passing through the Caribbean and the Straits of Florida. Thus the friction centered in the southwest is lessened and the Gulf Stream increases in volume, although the summer winds are comparatively weak. This seems to explain the double annual rhythm called for by the tide gauge record and by the computed transport values.

(8) Variations in transport having a period of more than one year probably depend both on the long period fluctuations in the strength of the general atmospheric circulation of the northern hemisphere and on the effectiveness of the northeast trade winds. Since the strength of the Gulf Stream System is the integrated result of all winds over the North Atlantic, it is supposed that mean sea level values along the southeastern coast of the United States, which reflect changes in the transport of the current, can be considered an index of the general atmospheric circulation.

(9) On the assumption that the clockwise current system of the North Atlantic surrounds a large core of relatively motionless water, the Sargasso Sea, gradual variations in the transport of the huge eddy can be expected under the circulation theorem to change its diameter and to cause fluctuations in the discharge of warm surface water towards northwestern Europe. Increasing currents should cause the eddy to contract and should lessen or even interrupt the flow of surface water towards the northeast. On the other hand, a long period of weakening currents should greatly increase the area of the northeastern North Atlantic covered by a relatively warm and saline surface layer. In this way quite small variations in the transport of the Gulf Stream may influence surface conditions at a considerable distance.

(10) While a prolonged decrease in the strength of the winds of the northern hemisphere by altering the climate can be expected eventually to lower sea surface temperatures in high latitudes, the first effect may be just the opposite, because of the necessity of decreasing the amount of warm water in the Sargasso Sea area. The converse, namely the temporary lowering of surface temperatures beyond the limits of the main eddy with a gradual strengthening of the wind system, is not so clearly indicated because the currents are free to make dynamic adjustment by contracting in diameter.

(11) There appears to be little possibility of securing adequate oceanographic surveys in the near future to test the validity of the assumption that the Gulf Stream System forms part of a relatively simple eddy. However, some preliminary correlation studies have indicated that it may be possible to link the Gulf Stream fluctuations with temperature changes in northern seas in such a way as to support the essential correctness of the much simplified circulation pattern assumed in this analysis.

(12) Perhaps one of the most significant results of this investigation is that the seasonal fluctuations in the transport of the Gulf Stream turn out to be greater than the year to year changes have been during the past 17 years. This means that in order to

learn about the effects of long-period variations in the volume of this current, it is only necessary to study the changes in current pattern and subsurface structure which occur each year in the North Atlantic. Man has often speculated concerning the meteorological and biological consequences of long-period changes in the strength of the Gulf Stream. Subsequent to 1922 the prolonged displacements of warm surface waters produced by such variations probably did not exceed those which might be observed in the course of any single year.

BIBLIOGRAPHY

DEITRICH, GUNTER

1937. Die Lage des Meeresoberfläche im Druckfeld von Ozean und Atmosphäre mit besonderer Berücksichtigung des westlichen Nordatlantischen Ozeans und des Golfs von Mexico. Veröffentlichungen des Instituts für Meereskunde an der Universität Berlin, Neue Folge, A, Heft 33, pp. 1–52.

EKMAN, V. WALFRID

1939. Ergebnisse und Probleme zur Theorie der Konvektionsströme im Meere. Ergebnisse der Kosmischen Physik, Band IV, pp. 1–74.

HACHEY, H. B.

1939. Temporary Migrations of Gulf Stream Water on the Atlantic Seaboard. Journal of the Fisheries Research Board of Canada, Vol. IV, No. 5, pp. 339–348.

ISELIN, C. O'D.

1926. A Study of the Circulation of the Western North Atlantic. Papers in Physical Oceanography and Meteorology, Vol. IV, No. 4, pp. 1–101.

1938a. Problems in the Oceanography of the North Atlantic. Nature, Vol. 141, No. 3574, pp. 772–776.

1938b. A Promising Theory Concerning the Causes and Results of Long-Period Variations in the Strength of the Gulf Stream System. Transaction of the American Geophysical Union, Nineteenth Annual Meeting, pp. pp. 243–244.

1939. Some Physical Factors Which May Influence the Productivity of New England's Coastal Waters. Journal of Marine Research, Vol. II, No. 1, pp. 74–85.

JAKHELLN, ANTON

1936. The Water Transport of Gradient Currents, Geofysiske Publikasjoner, Vol. XI, No. 11, pp. 1–14.

LAFOND, EUGENE C.

1939. Variations of Sea Level on the Pacific Coast of the United States. Journal of Marine Research, Vol. II, No. 1, pp. 17–29.

LEDANOIS, ED.

1934. Les Transgressions Oceanique. Off. Sci. Tech. Peches Maritimes, Revue des Travaus, Tome VII, Fasc. 4, No. 28, pp. 369–462.

MONTGOMERY, R. B.

1938. Fluctuations in Monthly Sea Level on Eastern U. S. Coast as Related to Dynamics of Western North Atlantic Ocean. Journal of Marine Research, Vol. I, No. 2, pp. 165–185.

ROSSBY, C.-G.

1939. Relation between Variations in the Intensity of the Zonal Circulation of the Atmosphere and the Displacements of the Semi-Permanent Centers of Action. Journal of Marine Research, Vol. II, No. 1, pp. 38–55.

SANDSTRÖM, J. W.

1903. Ueber die Anwendung von Pegelbeobachtungen zur Berechnung der Geschwindigkeit der Meeresstrome. Svenska Hydrografisk Biologiska Kommissionens Skrifter, Vol. I.

SEIWELL, H. R.

1937. Short Period Vertical Oscillations in the Western Basin of the North Atlantic. Papers in Physical Oceanography and Meteorology, Vol. V, No. 2, pp. 1–44.

1939. The Effect of Short Period Variations of Temperature and Salinity on Calculations in Dynamic Oceanography. Papers in Physical Oceanography and Meteorology, Vol. VII, No. 3, pp. 1–32.

SPILHAUS, A. F.

1940. A Detailed Study of the Surface Layers of the Ocean in the Neighborhood of the Gulf Stream with the Aid of Rapid Measuring Hydrographic Instruments. Journal of Marine Research, Vol. III, No. 1, pp. 51–75.

SVERDRUP, H. U.

1933. Vereinfachtes Verfahren zur Berechnung der Druck-Massenverteilung im Meere. Publikasjoner fra Chr. Michelsens Institutt, No. 26, Geofysiske Publikasjoner, Vol. X, No. 1, pp. 1–9.

1938. Research within Physical Oceanography and Submarine Geology at the Scripps Institution of Oceanography during April 1937 to April 1938. Transactions of the American Geophysical Union. Nineteenth Annual Meeting, pp. 238–242.

THOMPSON, E. F.

1939. A Rapid Method for the Determination of Dynamic Heights (or Depths) at Successive Lowerings at an Anchor Station. Journal of Marine Research, Vol. II, No. 2, pp. 126–130.