A number of papers in this issue of the JOURNAL OF PHYSICAL OCEANOGRAPHY were included in the program of the Conference on the Interaction of the Sea and the Atmosphere, which was held at Ft. Lauderdale, Fla., during the period 1–3 December 1971. These papers are identified by a footnote. It is anticipated that several more papers resulting from this conference will appear in forthcoming issues of this journal. Topics of the conference included: airborne observation techniques, macroscale air-sea interactions, air-sea materials exchange, boundary layer dynamics, and microscale air-sea interactions (see *Bulletin of the American Meteorological Society*, **52**, 921–940, for abstracts of all papers presented at this conference).

Negative Oceanic Heat Flux as a Cause of Water-Mass Formation[1,2]

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ABSTRACT

Regions of large (>60 kcal cm^{-2} year^{-1}) net annual heat flux from the oceans to the atmosphere are found only on the western sides of Northern Hemisphere oceans and in the Norwegian Sea, according to Budyko. It has been assumed that these negative heat fluxes are the result of the transport of warm water to middle and high latitudes by major ocean currents, specifically the Gulf Stream, the North Atlantic Current and the Kuroshio. In the case of the Norwegian Sea, Worthington suggested a mechanism for maintaining the negative heat flux, whereby warm surface water is advected northward to replace the dense, deep water formed within that sea by the cooling action of the atmosphere. The deep water mass so formed overflows the sills of the Norwegian Sea and moves southward into the depths of the Atlantic. It is postulated 1) that this mechanism maintains the negative heat fluxes found at mid-latitudes on the northwestern sides of the oceans, and 2) that water masses are also formed at these latitudes; they are intermediate rather than deep water masses. The water-mass producing agent in these northwestern regions of the oceans is polar continental air which breaks out over the sea surface at relatively low latitudes during northern winter.

1. Introduction

In his atlas of the heat balance of the earth's surface Budyko (1963) has presented, among others, a chart of the net annual heat exchange between the oceans and the atmosphere (Fig. 1). There are four regions in this chart where the net annual heat flux is strongly negative, i.e., regions where the oceans are yielding up heat to the atmosphere at a rate of more than 60 kcal cm^{-2} year^{-1}. Of these regions, three are found on the western sides of Northern Hemisphere oceans and the fourth in the Norwegian Sea.

Since a net annual heat flux from the ocean to the atmosphere cannot be maintained without a supply of warm water from lower latitudes, it has been generally assumed that these heat fluxes are the result of the northerly transport of the major ocean currents; Budyko specifically mentions the Kuroshio, the Gulf Stream, and the North Atlantic Current as being responsible for the strong negative heat fluxes shown in Fig. 1.

2. Negative heat flux in the Norwegian Sea

In the case of the Norwegian Sea it can be shown that this explanation is not wholly satisfactory. Since the work of Cooper (1955), it has become increasingly clear that the Norwegian Sea is the major source of North Atlantic Deep Water. The water mass is formed within that sea by the cooling action of the atmosphere which renders the water denser, at all depths, than the Atlantic water outside. In consequence it cascades out over the sills of the Norwegian Sea, between Greenland, Iceland and Scotland, and light surface Atlantic water is drawn into the Sea to takes its place. Accordingly, the negative heat flux over the Norwegian Sea is not merely the result of a warm inflow but is, in large part, the cause of the warm inflow.

Numerous observers (Steele *et al.*, 1962; Crease, 1965; Worthington and Volkmann, 1965; Hermann, 1967; Swallow and Worthington, 1969; Worthington, 1969) have made direct current measurements in the

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[2] Presented at the Conference on the Interaction of the Sea and the Atmosphere, 1–3 December 1971, Ft. Lauderdale, Fla.

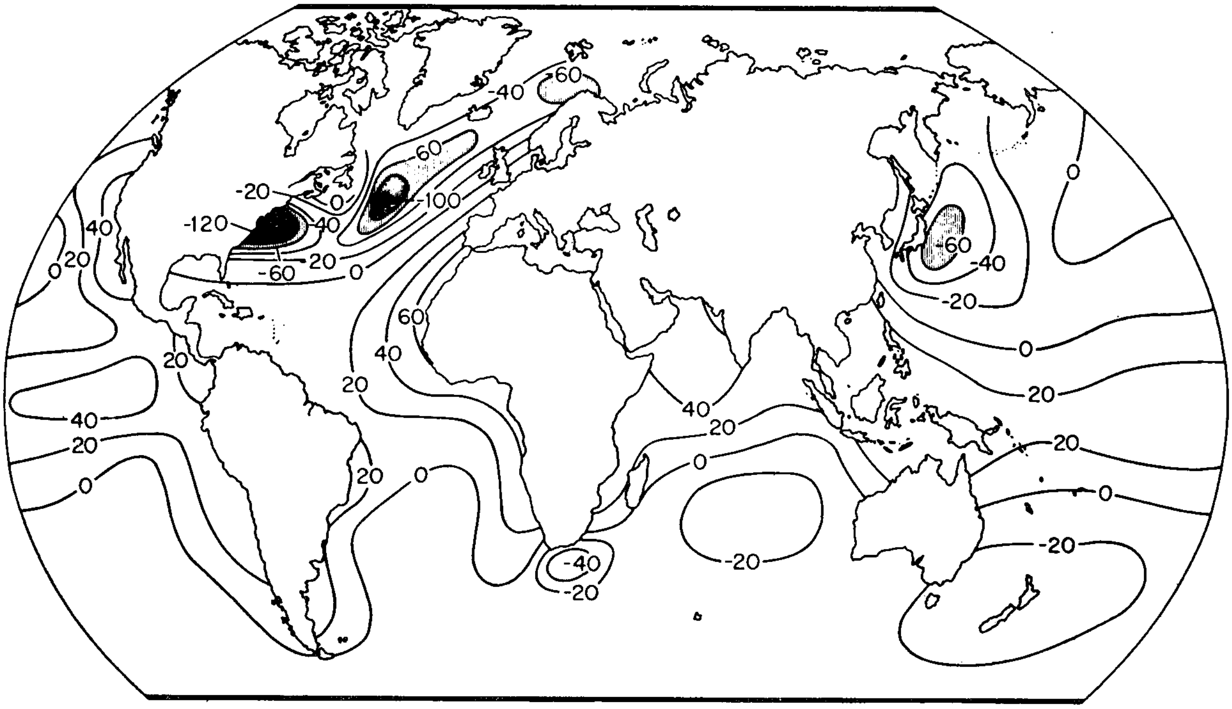

FIG. 1. Net annual heat flux between the ocean and the atmosphere, after Budyko (1963). Contours are in units of kcal cm^{-2} year^{-1}.

Norwegian Sea overflows, and there can be no doubt that they are active; in the last cited work, current velocities of as high as 140 cm sec^{-1} were reported at one current meter in the cold, overflow water passing through the Denmark Strait, and the loss of four neighboring moorings was described. This loss was apparently due to currents too strong to be withstood by the moorings. On the basis of all these foregoing measurements the volume transport of the overflows can be estimated at 6×10^6 m^3 sec^{-1} exclusive of entrained Atlantic water (4×10^6 m^3 sec^{-1}) which combines with the overflow water at the sills to bring the total production of North Atlantic Deep Water to 10×10^6 m^3 sec^{-1}.

In this case the negative heat flux in the Norwegian Sea can be clearly identified as the cause of the formation of a water mass. Worthington (1970) presented a water and heat budget for the Norwegian Sea (see Appendix). In it, the net annual heat flux to the atmosphere is 75 kcal cm^{-2} year^{-1} which, roughly speaking, is the amount of heat which must be withdrawn from the warm, surface Atlantic water entering the Norwegian Sea in order to convert it into cold, deep overflow water in sufficient quantity to match the overflow volume transports which have been measured in the Atlantic. This heat flux is somewhat larger than that of Budyko (1963) but there is no fundamental disagreement.

3. Negative heat flux south of New England

The other two regions of strong, negative heat flux in the Atlantic are found south of New England and east of Newfoundland (Fig. 1). In each of these regions the net annual heat flux to the atmosphere exceeds 100 kcal cm^{-2} year^{-1}, according to Budyko (1963). There is no doubt that the Gulf Stream does transport warm water from low latitudes into the region south of New England. Iselin (1936) noted that the Gulf Stream transported some water warmer than that found on either side of it, and he recognized that this was water brought from lower latitudes. This "warm core", as it is sometimes called, is confined to the upper layers; at all other depths the Gulf Stream represents the cool transition zone between the cold Slope Water, to the north, which has a shallow thermocline, and the warm Sargasso Sea, to the south, which has a deep thermocline.

During April 1960 eleven meridional oceanographic sections were made across the Gulf Stream between 68°30′W and 48°30′W, by Fuglister (1963). Fig. 2 shows three of these sections: the first at 68°30′W, the second at 58°30′W, and the third at 48°30′W. A fourth section, made in March 1964, runs in a northeasterly direction near 45W. Also shown is the path of the Gulf Stream during the period 10–20 April 1960 in which the sections were made. It can be seen from these sections that there is a very distinct warm core at 68°30′W; the surface of the Gulf Stream carries water warmer than 23C while the Sargasso Sea water to the south has a temperature of about 18C. In the second section the warm core has dwindled to a small area warmer than 18C (the maximum temperature was 18.4C); and in the third and fourth sections the warm core has vanished, and the water carried by the Gulf Stream is indistinguishable

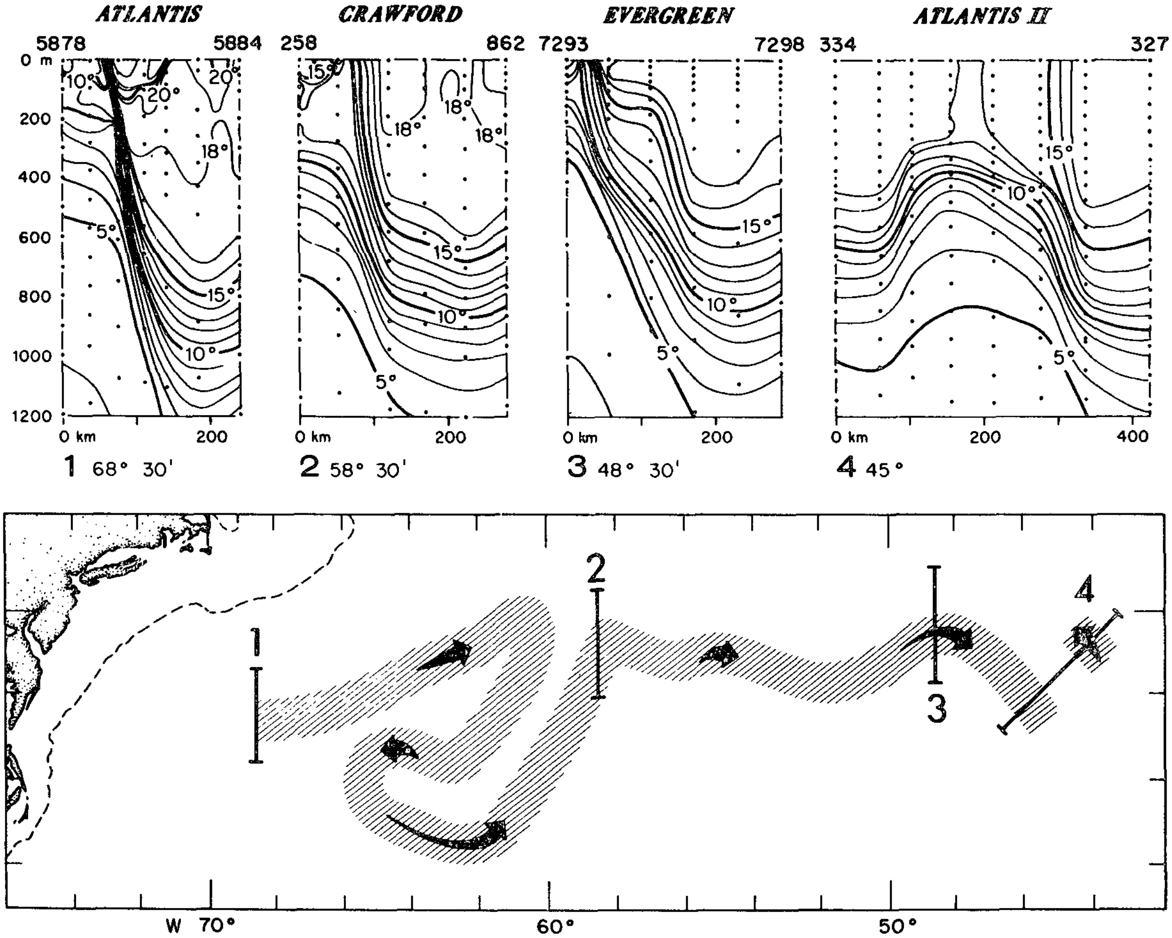

FIG. 2. Temperature sections (°C) across the Gulf Stream. Sections 1, 2 and 3 (April 1960) after Fuglister (1963); section 4 (March 1964), made by the author also crosses the North Atlantic Current (stations 331–334). Path of the Gulf Stream is shown as observed in April 1960.

from the Sargasso Sea water to the south. Clearly, the atmosphere is responsible for removing the major part of the warm core, but the agreement between the heat loss calculated from meteorological measurements and that calculated by geostrophic ocean heat transport is not as good as it was in the Norwegian Sea.

A rough calculation can be made of the amount of excess heat transported by the Gulf Stream from lower latitudes by assuming that all water with a temperature greater than 18C is of southern origin. Using Fuglister's (1963) volume transport calculations the amount of excess heat transport (above the 18C isotherm) for section 1 can be reckoned at 43.1×10^{12} cal sec^{-1}. Similarly, the amount of excess heat transport for section 2 where very little water above 18C was present was 0.5×10^{12} cal sec^{-1}. The warm core of the Gulf Stream has lost heat at the rate of 42.6×10^{12} cal sec^{-1} between sections 1 and 2 (Fig. 2). The surface area of the warm core can be calculated from Fig. 2; it was 1° of latitude (111 km) wide, and its length between sections 1 and 2, around the huge meander described by Fuglister (1963), was 2610 km. The area of major heat loss was thus 2.9×10^{15} cm^2. If the heat loss of the warm core was all to the atmosphere, it would have to be at a rate of 14.7×10^{-3} cal cm^{-2} sec^{-1}, or 1270 cal cm^{-2} day^{-1}. The evaporative heat flux calculated from meteorological data in the area during the 10-day period of the survey (10–20 April 1960) was between 730 cal cm^{-2} day^{-1} where the surface temperature was 23C, and 370 cal cm^{-2} day^{-1} where it had fallen to 18C.

It is clear, however, that all the loss of heat from the Gulf Stream was not due to the atmosphere: the transport of section 1 was much higher than that of Fuglister's (1963) other sections,[3] and the volume transport in the upper 100 m of the Gulf Stream diminshed 25% between sections 1 and 2. If this 25% transport loss is taken into account, the excess heat which should go to the atmosphere from the Gulf Stream is reduced from 1270 to 953 cal cm^{-2} day^{-1}, still well above the heat

[3] The very high transport of section 1 resulted in the discharge of warm water from the Gulf Stream to the Sargasso Sea to the south. This flow can be seen in Fuglister's (1963, p. 294) section at 66°30′W and was an unusual event. None of Iselin's (1940) thirteen sections in this region show such clear evidence of the warm core spreading southward, from the Gulf Stream.

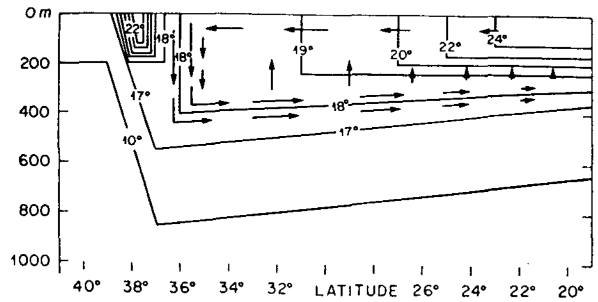

FIG. 3. Schematic temperature section (°C) and the meridional circulation induced by the formation of 18C water in the Sargasso Sea.

flux calculated from weather data. Regardless of the discrepancy between the meteorological and oceanographic methods of calculating heat loss, it is plain that on this occasion the low latitude water transported by the Gulf Stream has completely lost its excess heat, most of it to the atmosphere, by the time it reaches the 45th meridian. Other late winter sections in the northeastern Sargasso Sea such as the one shown here as section 4, made in March 1964, and those of Mann (1967), made in April 1963, confirm this.

South of the Gulf Stream (Fig. 2) a deep isothermal layer occurs in late winter. Worthington (1959) and Istoshin (1961) have suggested that in this region a water mass which they have termed 18C water is formed by the action of the atmosphere in late winter. This water mass can be seen as a nearly isothermal layer throughout the Sargasso Sea far to the south of its area of formation in latitudes where the winter minimum temperature remains well above 18C. It was first observed in April 1873 at 24°39′N by Thomson (1877) in the *Challenger*.

A schematic diagram (Fig. 3) shows the distribution of temperature in a north–south section across the Sargasso Sea in late winter. To the north, between 40–38N, is found the cold, slope water. The abrupt descent of the 17–10C isotherms takes place in crossing the Gulf Stream. The warm core can be seen in the upper 200 m where the temperature changes from less than 18C to more than 22C. Immediately to the south of the Gulf Stream a deep isothermal layer at 18C is found. A wide wedge of water between 17 and 19C extends to the south as far as the 20th parallel, and the main thermocline typified by the 10C isotherm (Iselin, 1936) shoals gradually from 850 to 650 m. The surface layer warms steadily toward the south. The 19–24C isotherms intersect the sea surface at the appropriate latitude for actual late winter conditions, and all the isotherms are drawn at the depths at which they occur in the real ocean. The arrows (Fig. 3) indicate the assumed meridional motion induced by the formation of 18C water. It is superimposed on the general zonal circulation, which can be deduced by the slopes of the isotherms; they indicate a narrow rapid eastward motion in the Gulf Stream and a slow wide return flow toward the west between the Gulf Stream and the 20th parallel.

The meridional circulation is assumed to be similar to that in the Norwegian Sea and the northern Atlantic. At high latitudes a dense water-mass is formed by the atmosphere, in this case by the outbreaks of cold, dry continental polar air which flow over the northern Sargasso Sea in late winter. This water-mass is formed in superabundance and flows off to the south at the 300 m level, and warm water is advected north to replace it. The 18C water circulation can thus be regarded as a model of the North Atlantic as a whole except that the permanent thermocline here represents the bottom. Otherwise, the main features are the same: there is a surface layer, a thermocline layer, and a relatively less stable bottom water. Usually, there is a similar oxygen minimum layer in the thermocline; in the majority of stations in the Sargasso Sea, the oxygen concentration at the core of the 18C water is higher than that of the water above it. An objection to the meridional circulation scheme presented in Fig. 3 can be raised on the ground that it requires some warming of 18C water north of 26N where Budyko has shown that the net annual heat flux is negative (Fig. 1). In order to budget heat strictly, the advection of water from south of 26N should probably be greater than indicated here (Fig. 3), and the vertical flow north of 26N should be less.

Estimates of the rate of production, if any, of 18C water would be possible if the net annual heat flux in the formation area were reliably known. In a recent study Warren (1972) has concluded that no production takes place; in his model the net annual heat flux at 35N, 60W, in the center of the production area assumed by Istoshin (1961), is zero, and the negative heat flux in winter balances the positive heat flux in summer and thus serves only to remove the seasonal thermocline. Examination of Budyko's (1963) chart of net annual heat flux (Fig. 1) gives a value of 50 kcal cm^{-2} $year^{-1}$ at this intersection. In a recent work Timonov *et al.* (1970) estimate the winter (October–March) heat flux south of the Gulf Stream to exceed 250 kcal cm^{-2} which is more than double the total for these months (92 kcal cm^{-2}) arrived at by inspection of Budyko's (1963) charts.

The area of formation of 18C water can be estimated at 14×10^{15} cm^2 (an area west of 45W, and between 33N and the Gulf Stream). If we take Budyko's annual heat flux of 50 kcal cm^{-2} $year^{-1}$ as an average, the heat given up to the atmosphere in this area would amount to 70×10^{19} cal $year^{-1}$. If we further estimate on the basis of Fig. 3 that the southern surface water loses an average of 3C en route to the formation area, the formation rate can be put at 23×10^{19} cm^3 $year^{-1}$, or 230×10^3 km^3 yr^{-1}. Wright and Worthington (1970) have provided a volumetric census of the water masses of the North Atlantic, and according to their worksheets the volume of 18C water in the Sargasso Sea is 890×10^3 km^3; thus, at this formation rate 26% of the 18C water would be renewed

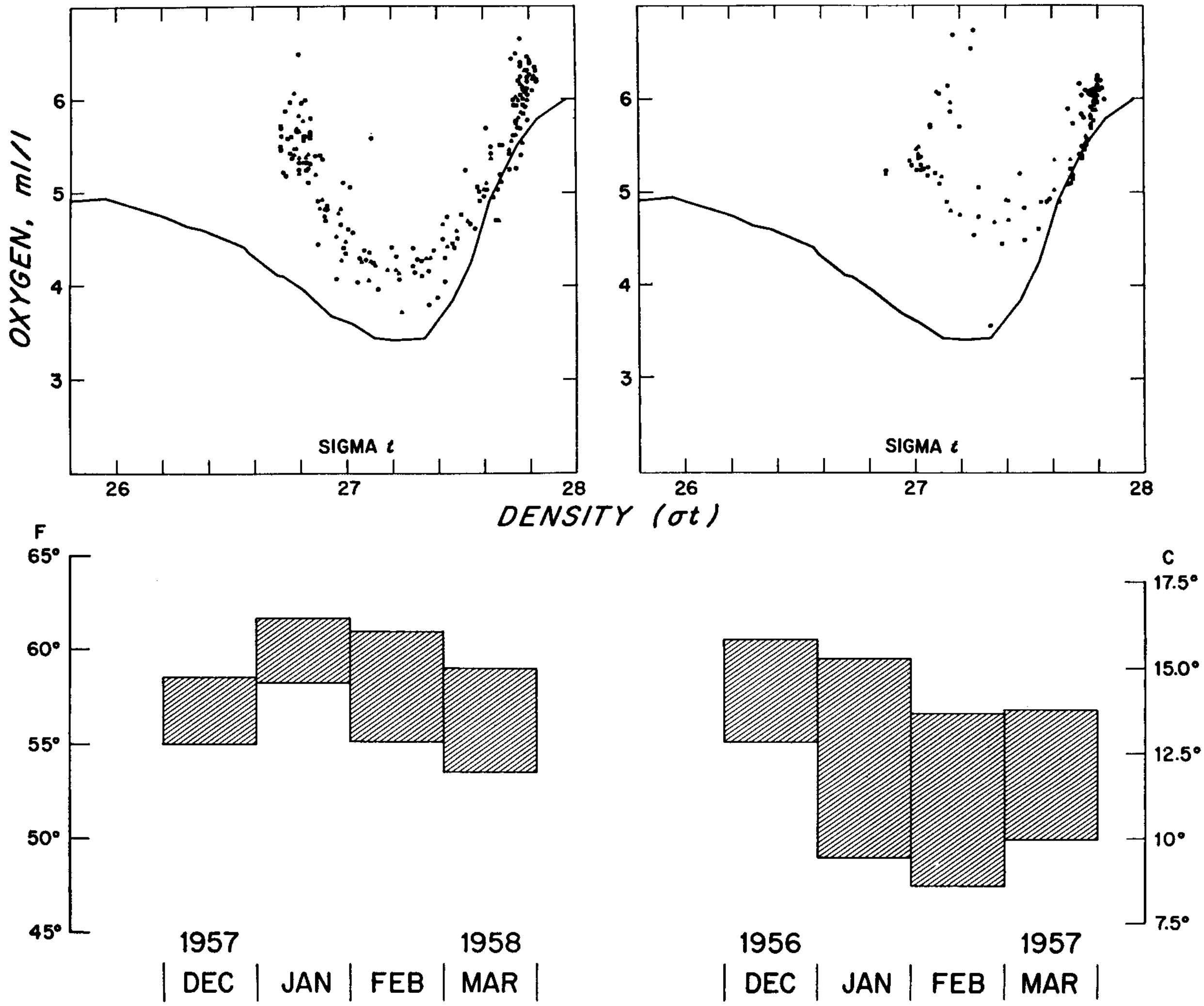

FIG. 4. Upper: Oxygen-density relationships for two sections across the North Atlantic Current: left, *Gauss* section, April 1958; right, *Discovery II* section, April 1957. Lower: Air-sea temperature difference at Weather Station Delta for the winter of 1957–58 (left) and 1956–57 (right).

each year. However, if we believe Timonov *et al.* (1970) this value would presumably be more than doubled, and if we believe Warren (1972) the value would be nil. Any confidence in estimates of this sort is clearly misplaced until the drag coefficient used in the calculation of evaporative heat flux is reliably known. Warren used 1.6×10^{-6} gm cm^{-3} while both Budyko and Timonov *et al.* used 2.5×10^{-6} gm cm^{-3} (with quite different results, presumably due to different meteorological data). It is of interest that Timonov *et al.* expressly state that the coefficient 2.5×10^{-6} gm cm^{-3} was "obtained by closing the thermal balance equation for the entire world ocean."[4]

4. Negative heat flux east of Newfoundland

However much 18C water is formed in the northern Sargasso Sea, one thing seems certain: it is not transported out of that sea to the northward by the Gulf Stream. In the temperature section at 45W (Fig. 2) water warmer than 17C can be seen flowing away from the observer (to the right) in a southeasterly direction in the top 400 m (the average temperature in the top 400 m for stations 326–328 is 17.5C and the average salinity 36.5‰). Water colder than 15C can be seen flowing toward the observer (to the left) in a northwesterly direction (0–400 m: $\bar{T}=14.5$C, $\bar{S}=36.0$‰). To convert surface Gulf Stream water (by now indistinguishable from Sargasso Sea water) at $T=17.5$C, $S=36.5$‰ into water of $T=14.5$C $S=36.0$‰ poses an impossible task to the atmosphere. Such a reduction in temperature would have to be accomplished in an area of relatively low heat flux (Fig. 1) and without benefit of any evaporative heat flux since the salinity must be reduced by 0.5‰. In the top 400 m, therefore, the discontinuous, two-gyre hypothesis of Worthington (1962)

[4] A coefficient of 2.04×10^{-6} gm cm^{-3} was used for the calculation of heat loss from the Gulf Stream in this paper.

is clearly preferable to continuous flow, and the northwesterly moving water between stations 331 and 334 is part of a separate anticyclonic gyre which is located beneath the second North Atlantic region of strongly negative net annual heat flux (Fig. 1) east of Newfoundland.

In this region the concentration of dissolved oxygen in the ocean pycnocline is generally higher than that in the Sargasso Sea. In the upper half of Fig. 4, dissolved oxygen is plotted against density (σt) for two sections which cross the North Atlantic Current east of Newfoundland near Weather Station Delta (44N, 41W). The solid line represents the oxygen-density relationship established by Richards and Redfield (1955) for the Sargasso Sea. In both these sections all the observed oxygen values in the pycnocline fall above the Sargasso Sea curve. Iselin (1939) first suggested that the thermocline water mass of the western North Atlantic was formed in late winter over a fairly wide range of latitude in that ocean. (The pycnocline and thermocline are virtually identical in the western North Atlantic.) The generally higher values of O_2 in the pycnocline east of Newfoundland tend to confirm Iselin's hypothesis, in that the density surfaces which form the pycnocline are in fact in direct contact with the sea surface between latitudes 45–55N in late winter. It must be admitted that the renewal process has never been observed directly, but Iselin (1939) assumed that renewal was accomplished by the sinking of surface water along density surfaces. It can be further assumed that such sinking is likely to be more active in regions where the heat flux is strongly negative than elsewhere.

In support of the latter assumption it can be seen (Fig. 4) that the O_2 concentration in the *Discovery II* section made in April 1957 was substantially higher throughout the pycnocline than it was in the *Gauss* section made in April 1958. In the lower half of Fig. 4 the monthly average air and water temperature for both years at Weather Station Delta are shown, with the shaded area representing the air-sea temperature difference. It can be seen that the winter of 1956–57 was more severe than the winter of 1957–58. The inference drawn here is that the renewal of the pycnocline is more active in a severe winter than in a mild winter.

It can thus be hypothesized, since the Gulf Stream does not contribute warm, low-latitude water to the region east of Newfoundland, that the negative heat flux found there is associated with the formation of the main thermocline water mass of the western North Atlantic and that warm water must be advected into the region from the southeast to replace it.

5. Conclusions

In summary, the hypothesis that regions of strong negative heat flux in the North Atlantic and Pacific are maintained solely by the transport of warm water to high latitudes by major ocean currents such as the Gulf Stream, the North Atlantic Current, and the Kuroshio is disputed. It is suggested that the effect of continentality in maintaining these negative heat fluxes has been neglected. As a result of continentality, the atmosphere delivers polar continental air to relatively low latitudes. This effect results in the formation of water masses by the sinking of surface waters on the western sides of Northern Hemisphere oceans. It is further suggested that the strong negative heat fluxes are maintained, in large part, by the advection of warm surface water from lower latitudes which flows northward to replace the water which has been converged and sunk in the formation process. It is implicit that such convergent processes can take place only at the end of winter after all the seasonal thermoclines in these regions have been erased.

It has been remarked elsewhere (Worthington, 1971) that oceanic anticyclones such as the Gulf Stream system, the Kuroshio circulation south of Japan, and the northern gyre of the North Atlantic east of Newfoundland, are also all situated on the western sides of Northern Hemisphere oceans, directly underneath the regions of strongly negative net annual heat fluxes. It has been suggested (*op. cit.*) that these anticyclonic current systems are in part the result of these excessive negative heat fluxes which, in turn, are the result of excessive continentality in the Northern Hemisphere. This hypothesis and the others presented in this paper have provided the author with a powerful incentive to measure the heat fluxes correctly.

Acknowledgments. I am indebted to P. M. Saunders of the Woods Hole Oceanographic Institution who performed the necessary heat flux calculations from the warm core of the Gulf Stream and to A. F. Bunker of the same Institution who read the manuscript critically and pointed out the inconsistency mentioned with regard to Fig. 3. This research was sponsored by the Office of Naval Research under Contract CO 241.

APPENDIX

Norwegian Sea Heat Budget[5]

Heat in[6]

Incoming water	Flow ($cm^3\ sec^{-1}$)	Temperature excess (°C)	Heat added ($cal\ sec^{-1}$)
Denmark Strait	1×10^{12}	6	6×10^{12}
Faroe-Shetland Channel	8×10^{12}	9	72×10^{12}
Total			78×10^{12}

Heat out[6]

Outgoing water	Flow ($cm^3\ sec^{-1}$)	Temperature deficit (°C)	Heat added ($cal\ sec^{-1}$)
North Polar Sea	3×10^{12}	3	9×10^{12}
Iceland-Scotland Ridge	2×10^{12}	1	2×10^{12}
Denmark Strait	4×10^{12}	1	4×10^{12}
Total			15×10^{12}

[5] Including the Barents Sea.

[6] The specific heat of sea water is taken to be 1 cal $(°C)^{-1}\ gm^{-1}$.

Heat loss to atmosphere

78×10^{12} cal sec$^{-1}-15\times10^{12}$ cal sec$^{-1}=63\times10^{12}$ cal sec^{-1}
Area of ice free Norwegian Sea $=2.64\times10^{16}$ cm^2

$$\therefore \quad \frac{63\times10^{12}\text{ cal sec}^{-1}}{2.64\times10^{16}\text{ cm}^2}\times3.15\times10^7\text{ sec year}^{-1}$$
$$=75\times10^3\text{ cal cm}^{-2}\text{ year}^{-1}$$

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