The Subpolar Mode Water of the North Atlantic Ocean

MICHAEL S. MCCARTNEY
Woods Hole Oceanographic Institution, Woods Hole, MA 02543

LYNNE D. TALLEY
WHOI/MIT Joint Program in Oceanography, Woods Hole, MA 02543

(Manuscript received 22 May 1980, in final form 8 July 1982)

ABSTRACT

The warm waters of the subtropical and subpolar basins of the North Atlantic have tight regional temperature-salinity relationships, and are conventionally called the regional "Central Waters." A volumetric census of the temperature-salinity characteristics of the North Atlantic by Wright and Worthington (1970) shows that waters characterized by certain segments of the *T-S* relationships have large volumes compared to those of other segments: volumetric "Mode Waters." Such Mode Waters appear as layers with increased vertical separation between isopycnals–pycnostads. The present study reports on the existence of pycnostads in the central and eastern North Atlantic. These Subpolar Mode Waters are formed by deep winter convection in the subpolar North Atlantic, and participate in the upper water circulation of the northern North Atlantic. The seasonal outcropping of the pycnostads occurs within and adjacent to the North Atlantic Current, the Irminger Current, the East and West Greenland Currents, and the Labrador Current. The warmer pycnostads (10°C $\leq T \leq$ 15°C) recirculate in an anticyclonic subtropical gyre east and south of the North Atlantic Current, causing volumetric modes in the central and eastern subtropical North Atlantic. A branch of the North Atlantic Current carries somewhat heavier and cooler (8°C $\leq T \leq$ 10°C) pycnostads northward past Ireland. The bulk of the current turns westward, but one branch continues northward, providing a warm core to the Norwegian Current (8°C). Within the main westward flow the density continues to increase and temperature to decrease. Southeast of Iceland pycnostad temperatures are near 8°C. Following the cyclonic circulation around the Irminger Sea west of the Reykjanes Ridge the temperature drops to less than 5°C. The cyclonic flow around the Labrador Sea gives a final pycnostad temperature below 3.5°C. The last, coldest, densest pycnostad is the Labrador Sea Water which influences lower latitudes via the southward flowing, Deep Western Boundary Current along the western boundary, and via eastward flow at mid-depth in the North Atlantic Current (Talley and McCartney, 1982).

1. Introduction

The notion that conditions at the sea surface at high latitudes play a major role in determining the temperature and salinity characteristics of the deep and thermocline waters at lower latitudes is an old idea. It evolved from a simple interpretation of the earliest measurements of low subsurface temperatures in the eighteenth century. Deep waters must somehow "originate" (acquire their coldness) at higher latitudes [Benjamin Thompson (Count Rumford), 1800]. In the 1930's a sharper image emerged. Wüst (1935) discussed a low-salinity layer in the lower thermocline of the South Atlantic Ocean and introduced the idea that the properties within individual layers of the thermocline are acquired in specific surface regions: "The vertical structure of the Subantarctic Intermediate Water, with its horizontal spreading at depths, is analogous to a vertical figure of the horizontal arrangement of temperature and salinity at the surface of the formation region." Montgomery (1938) provided a more formal interpretation using the meteorological concept of isentropic flow: isopycnal surfaces having a lateral (isopycnal) advective-lateral diffusive interior region. Non-isopycnal processes at edges of an isopycnal act as property source regions. In application to the tropical North Atlantic he found the sea-surface outcropping regions to be the dominant source regions for several isopycnals, with the isopycnal intersections with the sea floor and basin edges being secondary in importance.

The Montgomery hypothesis of sea-surface outcropping regions as isopycnal sources was further explored by Iselin (1939). The issue he discussed was the origin of the temperature-salinity relationship of the main thermocline in the North Atlantic—the "Central Atlantic Water" as he earlier named the relationship (Iselin, 1936). His comparisons of higher-latitude March sea-surface *T/S* data generally yielded a close correlation with the lower-latitude Central Water of the western and eastern subtropical thermocline. For example, in the temperature range

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6–11°C, the thermocline of the subtropical thermocline of the eastern Atlantic was shown to have a temperature-salinity relationship close to that of the sea-surface mixed layer in the subpolar latitudes of the Central Atlantic in winter. These ideas were extended to the rest of the world's oceans by Sverdrup *et al.* (1942), who found the correlation between the various regional Central Water relationships and the higher-latitude sea-surface properties to be a global phenomenon.

The aforementioned studies involved temperature-salinity relationships, with some supportive use of other properties like oxygen. There is another property that has been used to illustrate the influence of higher latitude outcroppings on the thermocline: the thickness of a water mass. In some regions of the open ocean the winter outcropping of isopycnals takes the form of deep convection. Deep convection produces a water mass characterized by vertical homogeneity, a pycnostad.[1] The homogeneity persists locally in time through the annual cycle and spatially along subsurface advection paths away from the region of convection. A familiar example is the Eighteen Degree Water of the western subtropical North Atlantic (Worthington, 1959, 1976).

The vertical and lateral homogeneity of the Eighteen Degree Water is reflected by the existence of a bivariate "mode" in a volumetric census of the temperature and salinity of the North Atlantic. Similar Mode Waters exist throughout the world's oceans (McCartney, 1982).

The present paper will focus on the contribution of deep winter convection in the subpolar North Atlantic Ocean to the Central Water of the North Atlantic Ocean. A surface circulation is inferred in which warm waters (>14°C) are modified to colder, fresher waters by heat and fresh-water exchange with the atmosphere along the circulation path. The progression of deep convective layers is the source of a variety of Subpolar Mode Waters ranging in temperature from greater than 14°C to less than 4°C. Here the *local* influence of convection on the volumetric characteristics will be shown, documenting the existence of the Subpolar Mode Water. In two related studies, the *remote* influence of subpolar convection on the thermocline and deep water at subtropical latitudes is described. Subtropical recirculation from the subtropical and subpolar convection regions to lower latitudes is illustrated by McCartney (1982) throughout the World Ocean. A more complete discussion of such recirculation in the North Atlantic is included in the present paper. The recirculation takes two forms. Lighter varieties of Subpolar Mode Water, with temperatures of 10 to 15°C, recirculate directly as part of the subtropical anticyclonic circulation, and contribute pycnostads to the Central Water of the subtropics. The densest variety of Subpolar Mode Water, the Labrador Sea Water, indirectly influences the Central Water at subtropical latitudes. The southward flow of the Deep Western Boundary Current includes the Labrador Sea Water which strongly influences the base of the subtropical main pycnocline—the densest part of the Central Water relationship. This influence is thoroughly described by Talley and McCartney (1982), using the low potential vorticity of the Labrador Sea Water as a tracer.

In a third study (McCartney and Talley, 1982) estimates of the conversion rate of warm to cold water north of 50°N are made.

[1] The suffix *stad* was introduced by Seitz (1967) as the antonym of *cline*. A thermostad is thus a layer of low vertical temperature gradient.

2. Large-volume warm water masses of the North Atlantic Ocean

The vertical homogeneity of a thick water mass and its existence over large geographic areas leads to a large volume contribution to the regional Central Water. This is illustrated in Fig. 1, which contains a simplified representation of the volume of water in the North Atlantic Ocean at temperatures above 4°C. The volumetric T/S diagram is constructed from the detailed census of temperature and salinity by Wright and Worthington (1970). Also included on Fig. 1 is the average temperature-salinity relationship for the North Atlantic Ocean (Wright and Worthington, 1970, Fig. 3). This relationship defines the Central Water of the North Atlantic Ocean as a whole. The Eighteen Degree Water is so thick that it shows as an isolated volumetric mode for the entire North Atlantic Ocean even though it is restricted geographically to the western subtropical North Atlantic. This modal character has led to the usage of the term "Mode Water" as a name for a component of a Central Water whose volume is largely due to formation by deep convection. In other words, a given Mode Water is a source water for a specific component of Central Water. Both Central Waters (Sverdrup *et al.*, 1942) and Mode Waters (Masuzawa, 1969; McCartney, 1977, 1982) are ubiquitous features of the world's oceans.

It is evident from Figure 1 that the Eighteen Degree Water is not the only large-volume water mass in the warm waters of the North Atlantic Ocean. There are other high-volume modes centered around the average T/S curve. [The two high-volume branches due to South Atlantic influence (fresh) and Mediterranean influence (saline) are discussed by Wright and Worthington (1970).] There are three major volumetric features along the average T/S curve:

1) The Eighteen Degree Water (17–18°C, 36.4–36.5‰).

2) A modal ridge in the temperature range 10–

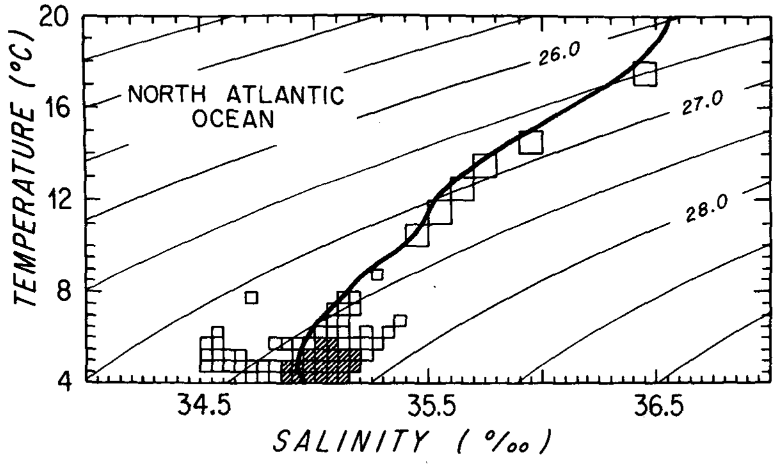

FIG. 1. Simplified representation of the Wright and Worthington (1970) volumetric census of the temperature-salinity characteristics of the North Atlantic Ocean for the warm water ($T \geq 4°C$). The various T-S classes are first ranked by their volume per 0.5°C and 0.05‰. The highest-ranked classes that collectively total 25% of the total warm water volume are represented by the hatched boxes. The highest-ranked classes that collectively total 50% of the total warm water volume are represented by the combination of hatched and unhatched boxes. The heavy curve represents the average temperature-salinity relationship for the North Atlantic Ocean. After Wright and Worthington (1970).

15°C, with maximum volume at 11–12°C, 35.5–35.6‰ (Subpolar Mode Water).

3) High-volume, cold modes below 6°C, with maximum volume at 4–4.5°C, 34.95–35.0‰ (Subpolar Mode Water and Labrador Sea Water). These are adjacent to and warmer than the even-larger-volume North Atlantic Deep Water (defined as colder than 4°C by Wright and Worthington, 1970).

The second and third features are the focus of the present study. The Eighteen Degree Water has been thoroughly discussed in other papers and is not directly connected with the Subpolar Mode Water. The reader is particularly referred to Worthington (1959, 1976) for further characteristics of the Eighteen Degree Water, to Leetmaa (1977) for a description of a convective formation event, to Jenkins (1982) and Talley and Raymer (1982) for description of decadal variability in its characteristics, and to McCartney (1982) for a description of its potential vorticity characteristics.

In addition to their census of the North Atlantic Ocean as a whole, Wright and Worthington (1970) did regional censuses in six basins of the Ocean. The boundaries of these basins (Fig. 2) were picked on the basis of amounts of high-quality data available, and of deep-water property distributions and topographic features. For the present study the two subpolar basins, the Labrador and the European, and the two subtropical basins, the North American and the North African, are the most relevant. The volumetric T/S relationships for these four basins are shown in Fig. 3 from the Wright-Worthington (1970) atlas. The "Central Waters" in each basin can be defined as the basin-averaged T/S relationships shown in the figure; these curves are also from the Wright-Worthington (1970) atlas. For example, the North Atlantic Basin relationship is quite close to the Iselin (1936, Fig. 53) Central Atlantic Water curve defined from Sargasso Sea data.

The large-volume temperature-salinity classes in each basin are not uniformly clustered along the Cen-

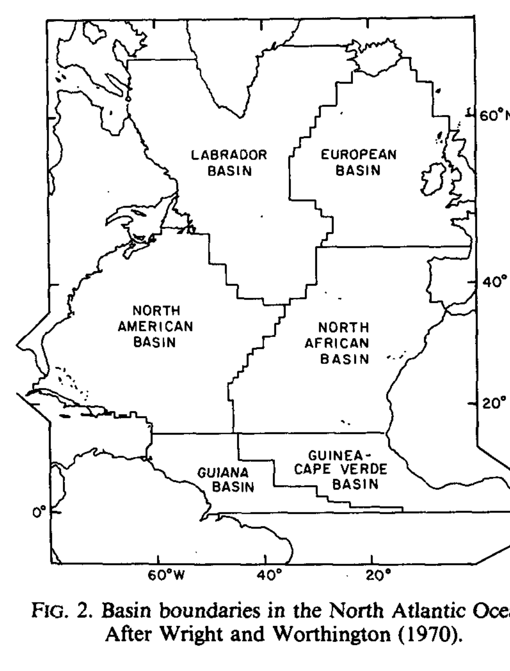

FIG. 2. Basin boundaries in the North Atlantic Ocean. After Wright and Worthington (1970).

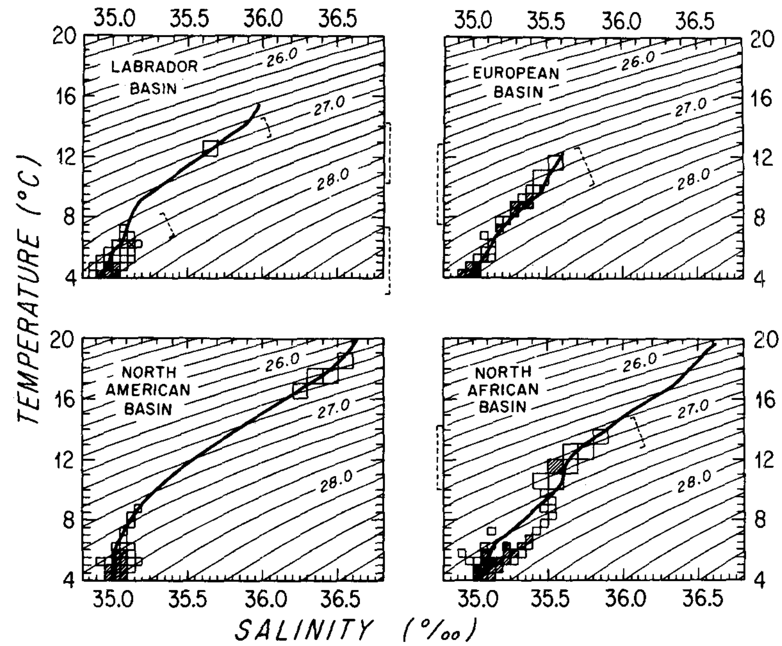

FIG. 3. As in Fig. 1, but constructed from the Wright and Worthington (1970) individual basin (Fig. 2) census charts and temperature-salinity relationships. The dashed brackets indicate the temperature and density ranges for deep convection within the two northern basins, and for subtropical recirculation in the North African Basin (see text). Note the single enclosed *T-S* class indicated by an × in the Labrador Basin chart. This box is not included in those adding up to 50% of the warm water.

tral Water relationship. Volumetric modes are found in each of the basins. The North Atlantic Basin mode in the 16–19°C temperature range is the Eighteen Degree Water. In the North African Basin, the largest volume is at 11–12°C. In the European Basin the temperature of the highest-volume water ranges from 7 to 12°C, separated by lower-volume classes from the large-volume classes below 5°C. In the Labrador Basin most of the water is colder than 5°C, but there is also a volumetric mode at a much higher temperature, 12–13°C.

The Eighteen Degree Water, or Subtropical Mode Water as it is sometimes called (Masuzawa, 1969; Warren, 1972), is the product of deep winter convection. In the next section, it is shown that the large-volume modes described above in the discussion of Fig. 3 are also of convective origin, and constitute a Subpolar Mode Water.

3. Deep winter convection in the North Atlantic and the Subpolar Mode Water

The influence of deep winter convection in the northern North Atlantic Ocean is easily recognized in the many temperature-salinity sections published in physical oceanographic atlases. In non-winter sections, for example most of those in the Fuglister (1960) atlas, there are thermostads and halostads in the upper water, generally at depths between 200 and 1000 m. These appear to be remnants of the winter mixed layer, locally isolated from the sea surface by a seasonal pycnocline. Winter sections are included in the atlases of Mann *et al.* (1965), Grant (1968), Dietrich (1969) and Worthington and Wright (1970). These show that over large areas of the northern North Atlantic, the near-surface pycnocline disappears at the end of winter and homogenization of the upper water column penetrates to hundreds of meters. A major product of the present study is the synthesis of winter data from various sources in charts of the properties of this deep, homogenized layer.

The seasonal change in upper-ocean density is illustrated in Fig. 4. This example comes from the Rockall Channel, in the European Basin west of Ireland. In late summer (Fig. 4a) there is a pycnostad sandwiched between the main pycnocline (below 800 m) and the seasonal pycnocline (above 200 m). The buoyancy period is used to measure the vertical homogeneity: the most homogeneous water has the highest buoyancy period. The most homogeneous water in this profile is designated the "pycnostad core" and is at a potential density of σ_θ = 27.36 mg cm^{-3}, centered at 400 m. The temperature-salinity characteristics of this pycnostad fall on the high-volume ridge for the European Basin (Fig. 3). The large volume of water along this "ridge" is due to the thickness of such pycnostads and their existence over a large area of the Basin.

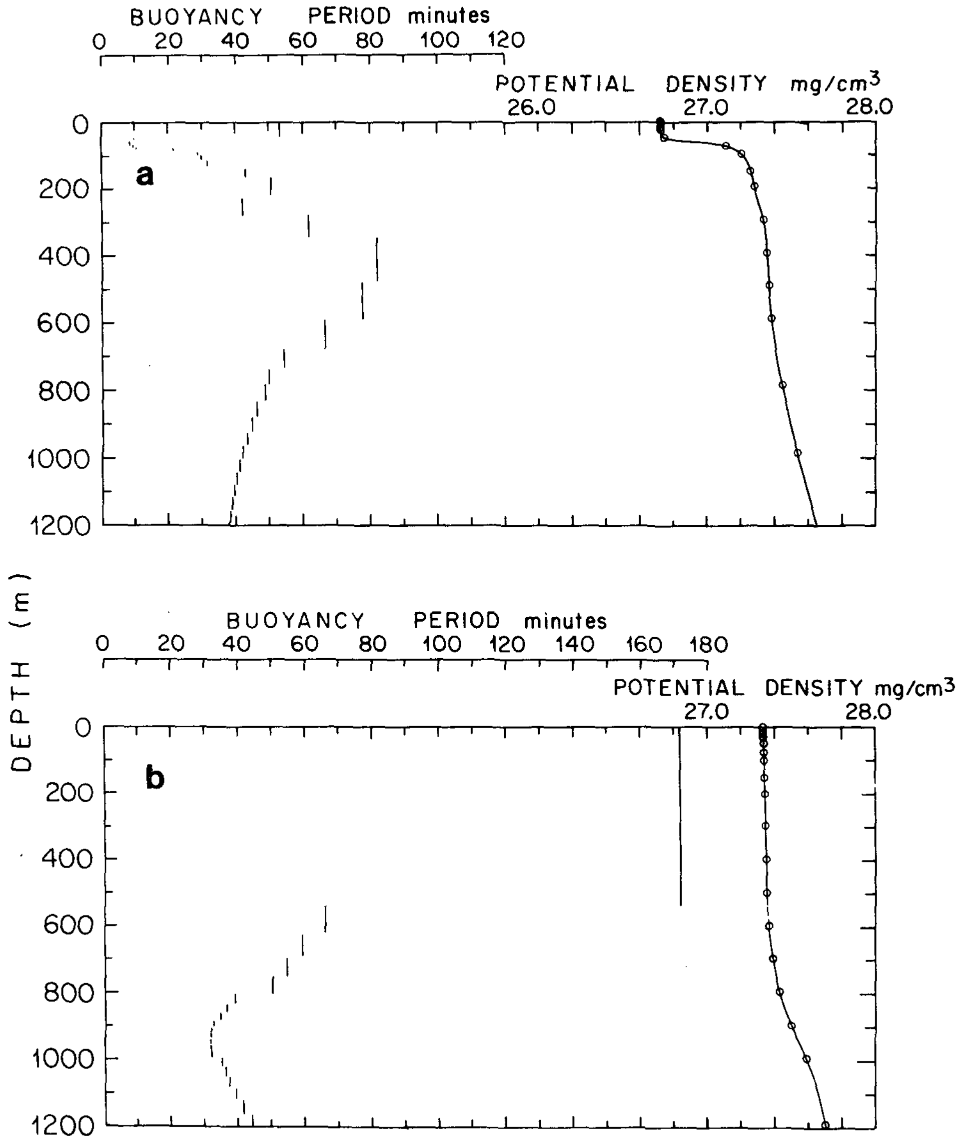

FIG. 4. Seasonal contrast of profiles of potential density and buoyancy period as functions of depth in the Rockall Channel: (a) H.M.S. *Dalrymple*, Station 24, occupied at 56°14′N, 9°38′W in late summer, 17 September 1963. The pycnostad (period maximum) centered near 400 m has $\theta \approx 9.5°C$ and $S \approx 35.39‰$. (b) H.M.S. *Dalrymple*, Station 6, occupied at 56°17′N, 9°48′W in late winter, 15 March 1964, six months after the station of (a). The convective mixed layer has $\theta \approx 9.8°C$ and $S \approx 35.42‰$.

A late winter station (Fig. 4b) made six months later and at nearly the same location, shows that the shallow pycnocline capping the pycnostad during the warm season disappears after the cooling season. The pycnostad is now open to the atmosphere and extends down to 540 m.[2] The temperature and salinity of the outcropping pycnostad differ only slightly from those of the preceding summer's pycnostad (it is warmer by 0.3°C and saltier by 0.03‰). The major change is in the vertical homogeneity: there is more than a four-fold increase in the thickness of a layer defined by a potential density interval of 0.02 mg cm^{-3} and more than a two-fold increase in buoyancy period.

A northern North Atlantic data set that will be referred to as "late winter" is illustrated in Fig. 5a. The cruises are listed in Table A1 in the appendix, along with various references containing vertical sections from the indicated cruises. "Late winter" is defined as January–April, the rather broad four-month period being necessary to achieve adequate spatial coverage. This could cause difficulties in producing a chart of the properties of the fully developed, deep

[2] The mixed layer depth is defined in this study by the depth over which the potential density increases by $\Delta\sigma_\theta = .02$ mg cm^{-3}.

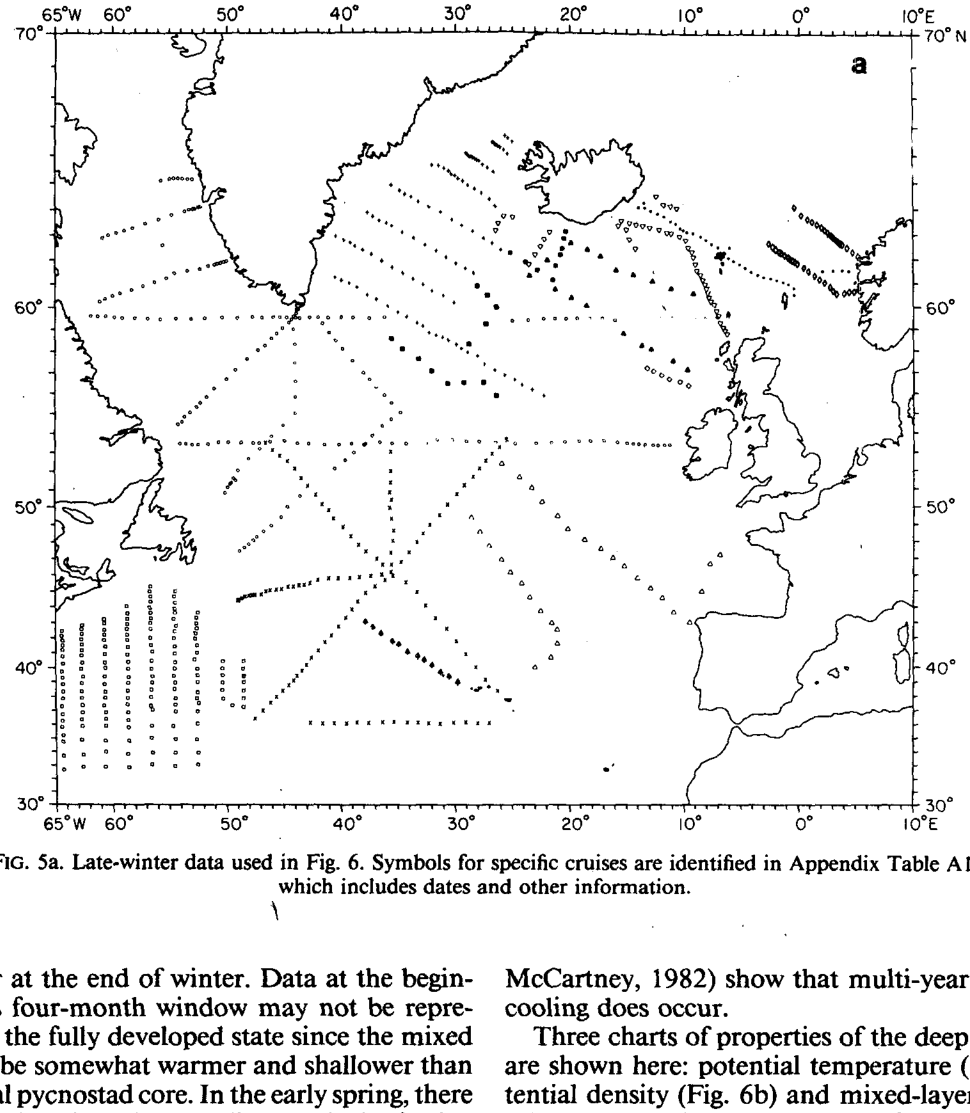

FIG. 5a. Late-winter data used in Fig. 6. Symbols for specific cruises are identified in Appendix Table A1, which includes dates and other information.

mixed layer at the end of winter. Data at the beginning of this four-month window may not be representative of the fully developed state since the mixed layer could be somewhat warmer and shallower than the true local pycnostad core. In the early spring, there may be mid-day diurnal pycnoclines or the beginning of the full seasonal pycnocline, and the recently homogenized pycnostad may be slightly isolated from the sea surface. Such caps have been ignored if they amount to a superficial potential density lid of magnitude less than 0.1 mg cm^{-3}.

Interannual variability presents an additional problem. The severity of a winter may vary from year to year and place to place, so that late-winter conditions in a particular year may be quite different from the long-term mean, late-winter condition. The data set described in the appendix Table A1 includes five different years in the nine-year period 1958–67. In addition to the concern about consistency between any given year and the long-term mean, there may be long-term trends to confuse the mapping further. Reports of long-term trends in Eighteen Degree Water (Jenkins, 1982; Talley and Raymer, 1982) and Labrador Sea Water (Lazier, 1981; Talley and McCartney, 1982) show that multi-year warming or cooling does occur.

Three charts of properties of the deep mixed layer are shown here: potential temperature (Fig. 6a), potential density (Fig. 6b) and mixed-layer depth (Fig. 6c). "Deep" mixed layers are defined to be thicker than 200 m. Considering the possible sampling difficulties mentioned above, the charts were remarkably simple to contour, the few main difficulties arising where fronts and cruise tracks overlapped and induced artificially large gradients, or data inconsistencies. In some of these regions segments of sections have been deleted; for example, the *Erika Dan* 59°N section between 22–36°W.

Common to these three charts is a heavily shaded area where the mixed layer is less than 200 m thick. The 200 m cutoff is admittedly arbitrary. In the northern North Atlantic, as noted at the beginning of this section, pycnostads generally are centered *below* 200 m. It appears that the seasonal thermocline obliterates the effects of shallower convection. Shallower convection may persist as a weak pycnostad, but is not likely to contribute much volume to the regional Central Water. In any case, there is no clear evidence

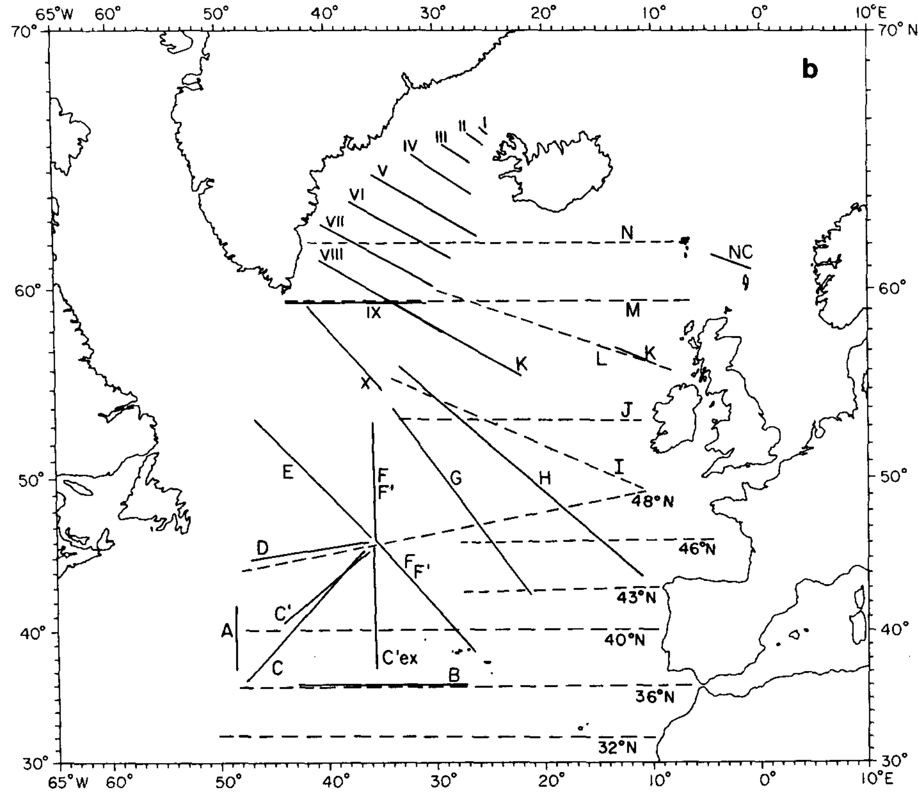

FIG. 5b. Location of sections used in Figs. 8, 9, 10, 11, 14, 15, 16 and 17. Details of stations used are identified in Appendix Table A2.

of annual persistence of pycnostads from shallower water winter convection in the various atlases cited earlier.

Both Subtropical and Subpolar Mode Waters are evident in the distribution charts. Part of the North American Basin is included in the southwest. Here the Eighteen Degree Water outcrops with temperature near 17.5°C and density near $\sigma_\theta \approx 26.55$ mg cm^{-3}. In the rest of the North Atlantic, the deep mixed layer is the Subpolar Mode Water. The outcropping Eighteen Degree Water is separated from the Subpolar Mode Water by an area of shallow convection overlying the Southeast Newfoundland Rise. The separation is defined by a single station near 46°W, 38°N, at the controversial "trough" between the two anticyclonic gyres in Worthington's (1976) monograph. It appears from an analysis of several data sets in this region that the outcropping and circulation of the Eighteen Degree Water is physically separated from the outcropping and circulation of the Subpolar Mode Water northeast of the Southeast Newfoundland Rise.

Over the rest of the subpolar north-central and northeastern North Atlantic, the deep convection is arranged in a basin-sized annular distribution. This continuous band of deep convection is broken in only one region: on the western boundary near 50°N where the North Atlantic Current separates from the boundary and penetrates eastward across the North Atlantic near 52°N. South of the separation point the warmest and lightest deep mixed layers are found, northwest of it the coldest and densest layers are found in the Labrador Sea. Fig. 7 shows examples of the density profiles from these two regions. The mixed layers can be quite deep, exceeding 1000 m. Wrapped around the eastward protrusion of the North Atlantic Current is a deep convection band with counterclockwise trends of decreasing temperature (smooth trend), increasing density (smooth trend) and increasing depth (irregular trend).

The volumes of water implied by the mixed-layer thicknesses and the surface areas they occupy are considerable. Consider for example the mixed layers between 11 and 12°C in Fig. 6b. The associated surface area is approximately 10^6 km^2. This surface area is only 2.7% of the total North Atlantic Ocean surface area. The average thickness of the 11–12°C layer in the North Atlantic is about 64 m. However, in the outcropping region the thickness is about five times this average, on the order of 300 m. Despite the small outcropping area of the 11–12°C water, the volume of 300×10^3 km^3 of 11–12°C water in the winter

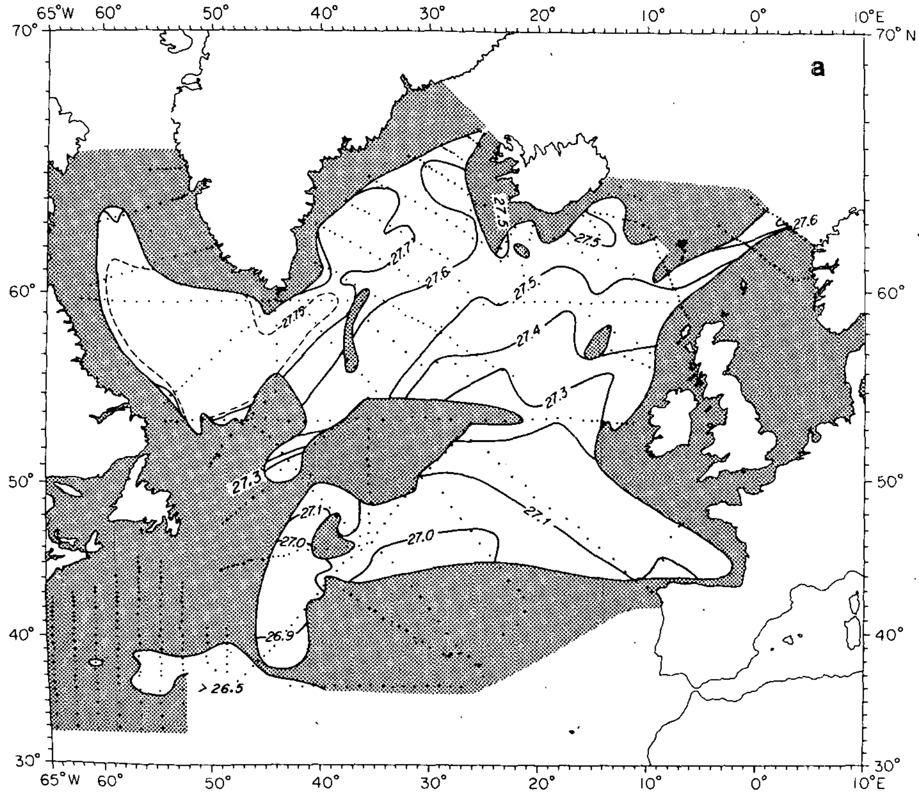

FIG. 6a. Potential density σ_θ (mg cm^{-3}) of the surface mixed layer in "late winter"—January through April. Properties are contoured only where the layer thicknesses are greater than 200 m for a density layer of $\Delta\sigma_\theta$ = 0.02 mg cm^{-3}. The shaded area occurs where the layer thickness is less than 200 m. The Norwegian-Greenland Sea is excluded from the contouring, except where shown in the Norwegian Current.

mixed layer is over half of the total volume of 11–12°C water (531 × 10^3 km^3) in the combined Labrador and European Basins (Wright and Worthington, 1970). The volume of mixed-layer water is 12.8% of the total 11–12°C water in the entire North Atlantic Ocean, 2348.6 × 10^3 km^3.

There is good correspondence of mixed-layer properties in Fig. 6 and the properties of the high-volume Mode Waters in Fig. 3. The *T/S* diagrams for the two subpolar basins in Fig. 3 include the density ranges and temperature ranges of deep convection within that basin. These ranges were obtained by simply overlaying the Basin boundaries from Fig. 2 on the data maps used to produce Figs. 6a and 6b.

In the North African Basin diagram of Fig. 3, recirculation from the convection zone farther north gives the relatively high volumes between 15 and 10°C. The southern edge of the deep-mixed-layer region barely penetrates south of the European and Labrador Basins into the North African Basin. The temperature and density ranges from the southern part of the pair of subpolar basins have been included on the North African Basin *T/S* diagram in Fig. 3. This part of the convection zone might be expected to communicate with lower latitudes as part of the subtropical anticyclonic recirculation of the North Atlantic Current. The evidence of such recirculation, briefly described by McCartney (1982), is the existence of Subpolar Mode Water at, say, 36°N, lying well beneath the sea surface but above the Mediterranean water. A relative volumetric minimum falls between 8.5 to 10°C, marking the center of the transition between the overlying recirculating Subpolar Mode Water and the underlying Mediterranean water. In Section 4a we provide more complete documentation of this southward recirculation.

4. Circulation in the northern North Atlantic

The North Atlantic Current is a major element of the circulation of the northern North Atlantic. Iselin (1936) specified a terminology for the various northerly and easterly currents at mid-latitudes in the North Atlantic. His ocean-wide "Gulf Stream System" is composed of, from southwest to northeast, the Florida Current, the Gulf Stream and the North Atlantic Current. The last of these names applies to "all the easterly and northerly currents from the point east of the 'tail' of the Grand Banks, where the Gulf Stream first divides." He noted that the subsurface paths of these currents are often masked by a shallow, wind-driven, superficial layer, the North Atlantic Drift.

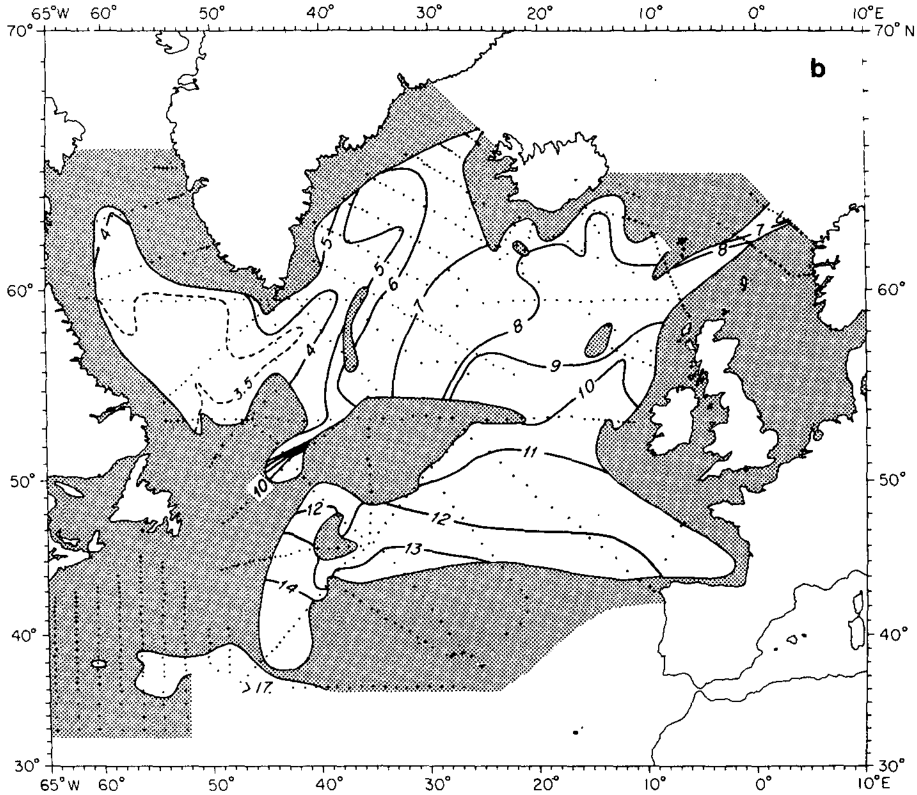

FIG. 6b. As in Fig. 6a except for potential temperature θ (°C).

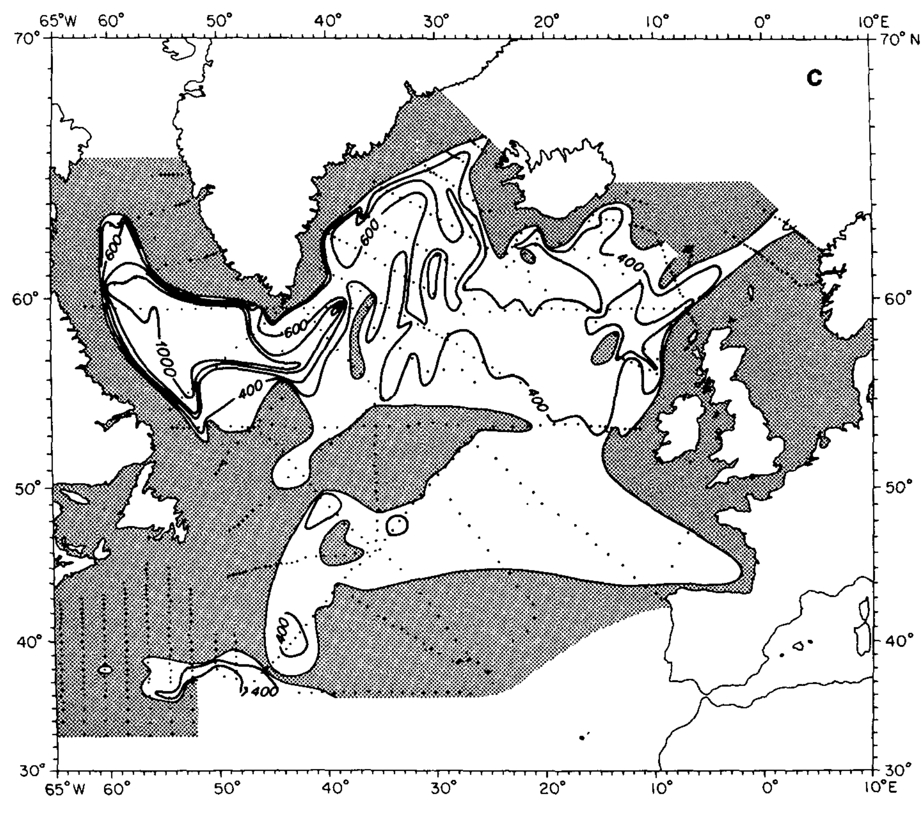

FIG. 6c. As in Fig. 6a except for mixed layer depth—thickness of a density layer of $\Delta\sigma_\theta = 0.02$ mg cm^{-3} at or close to sea surface.

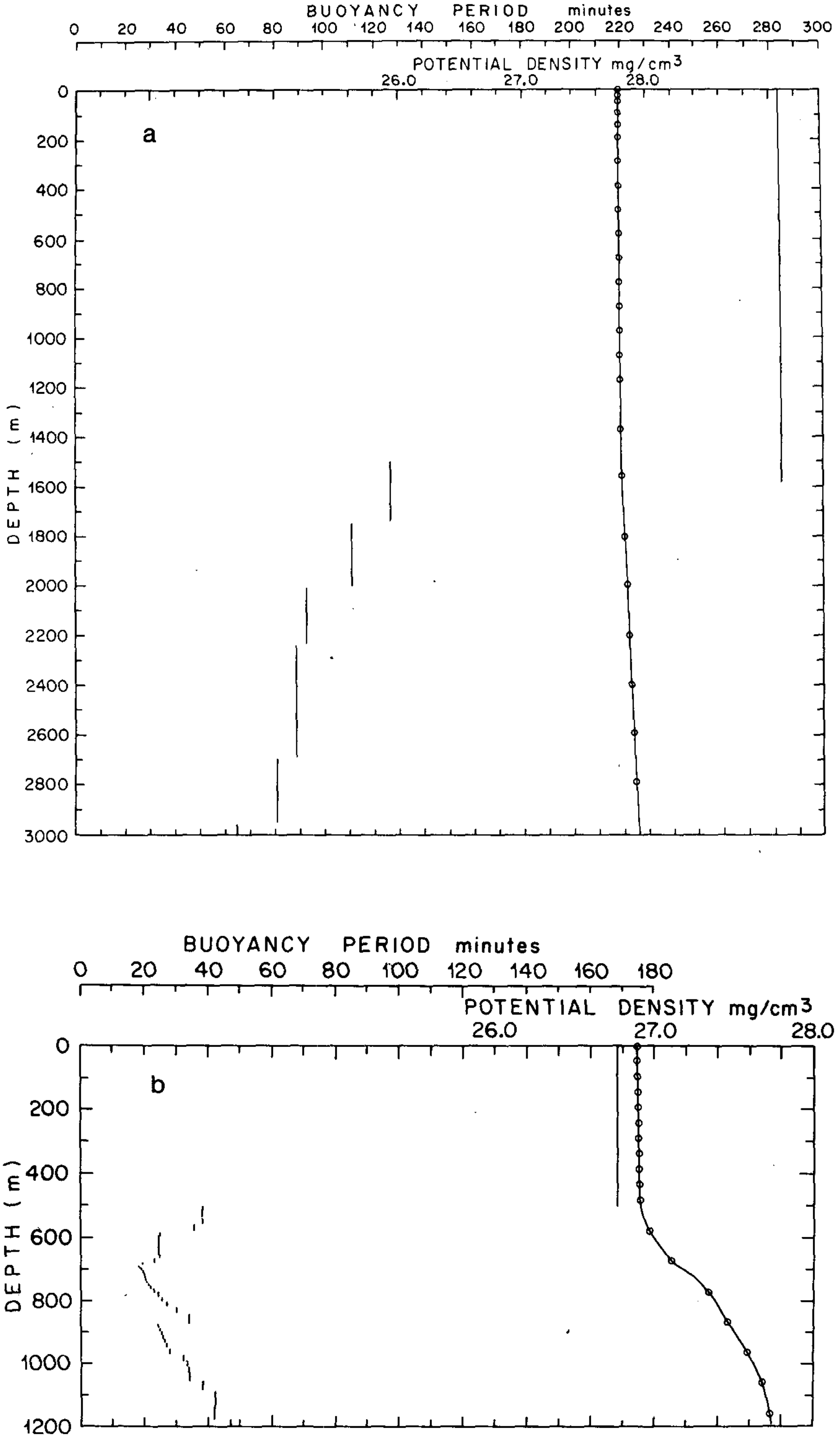

FIG. 7. Spatial contrast of profiles of potential density and buoyancy period as functions of depth between the regions of most extreme type of Subpolar Mode Water. (a) Labrador Sea, where densest varieties of Subpolar Mode Water outcrop. M.S. *Erika Dan* station 237, occupied at 57°37'N, 48°08'W in late winter, 26 February 1962. The convective mixed layer has $\theta \approx 3.5°C$ and $S \approx 34.88‰$. (b) North Atlantic Current near the Tail of the Grand Banks where the lightest varieties of Subpolar Mode Water outcrop. The convective mixed layer has $\theta \approx 14.7°C$ and $S \approx 36.08‰$. R.V. *Atlantis II* station 336, occupied at 40°45'N, 42°33'W in late winter, 18 March 1964, about 1100 n mi south-southeast from the station in (a), and two years later. It is 1600 n mi southwest of the station in Fig. 4b, and three days later.

The general circulation of the northern North Atlantic has two regimes, an anticyclonic one and a cyclonic one. The North Atlantic Current is the western boundary current for the northern part of the anticyclonic regime. This Current leaves the boundary near 52°W and extends eastward across the ocean, forming a boundary between the two regimes. South of this extension is the anticyclonic circulation, while north of the extension the circulation is cyclonic. A familiar illustration of this basic pattern is given by Sverdrup *et al.* (1942, Fig. 187).

In this section the relation of the North Atlantic Subpolar Mode Water to the North Atlantic Current and the anticyclonic and cyclonic circulations is discussed. The main tool in this discussion is a set of density sections crossing the Current (Fig. 5b). The relationship of the Mode Water to the anticyclonic regime is discussed in Section 4a. This anticyclonic regime recirculates part of the North Atlantic Current transport, but not all of it. The remainder branches northward, functioning as a source of warm water for the cyclonic circulation. This is described in Section 4b.

a. Transport of Mode Water by the North Atlantic Current and its subtropical, anticyclonic recirculation

South of the Grand Banks of Newfoundland, the Gulf Stream flows eastward, transporting Eighteen Degree Water. The Stream (Worthington, 1976) or a branch of the Stream turns southeast, and with it all the Eighteen Degree Water. This turning is illustrated by the progression of sections A, C and B in Fig. 8. The Eighteen Degree Water is a thick layer at $\sigma_\theta \approx 26.55$ adjacent to and overlying the tilted pycnocline that indicates the Stream. It appears from the distribution of Eighteen Degree Water (Worthington, 1976) that it recirculates back to the west, and does not contribute to the northward flow east of the Grand Banks.

East of the Grand Banks of Newfoundland, the North Atlantic Current flows northward with a strong signature in the density field. The deep mixed layers of the local Mode Waters are found both in the offshore part of the Current,[3] where the pycnocline descends to its maximum depth and farther offshore where the pycnocline slope reverses. Sections C and D in Fig. 8 show the northward flow of the North Atlantic Current with overlying deep mixed layers of the Subpolar Mode Water. Worthington (1976, Table 3) calculated the Current transports for sections C and D, with the sea floor as a zero-velocity surface. About 20 or 30% of the total transport occurs in the temperature classes containing the individual sections' deep mixed layers. If transports were computed relative to a mid-depth reference level, these percentages would be even larger. The local Mode Water is thus a major component of the northward transport of the North Atlantic Current in the region east of the Grand Banks.

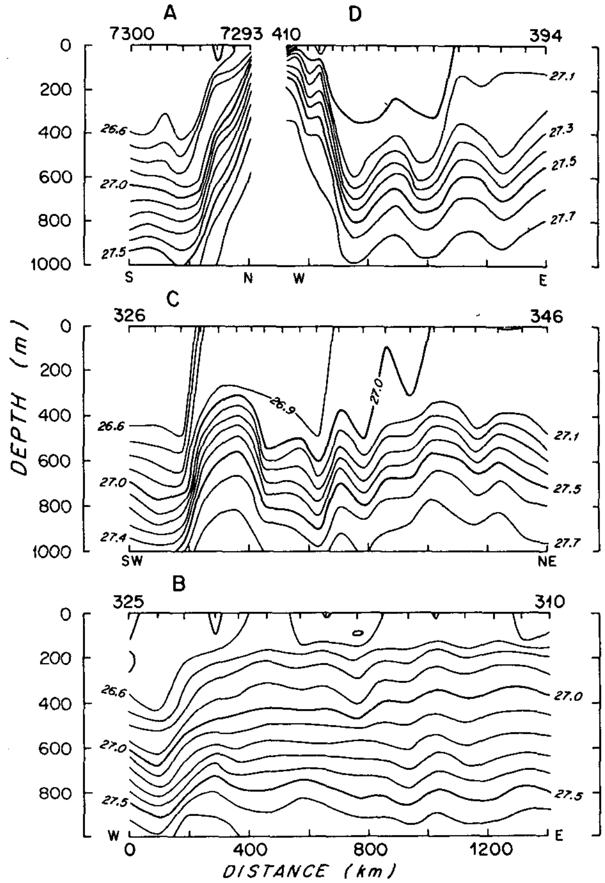

FIG. 8. Potential density for sections A–D (Fig. 5b and Appendix Table A2). Sections A, C and B include the eastward (A), southeastward (C) and southward (B) transport of the Gulf Stream, with the associated Eighteen Degree Water appearing as a thick layer at $\sigma_\theta \leq 26.6$ mg cm^{-3}. Sections C and D include the northwestward (C) and north-northwestward (D) transport of the North Atlantic Current, with the associated Subpolar Mode Water appearing as a thick layer at densities in the range 26.8 mg cm^{-3} $\leq \sigma_\theta \leq$ 27.1 mg cm^{-3}. In sections C and D, offshore of the North Atlantic Current, the pycnocline slope reverses, and the southward flowing Mode Water is somewhat denser than that in the Current.

The Current turns offshore near 52°N in this particular data set. Fig. 9 shows four additional sections; the first, E, from the neighborhood of this separation and the other three, F, G and H, from farther east. The association of deep mixed layers with the north-to-south descent of the pycnocline shows that the Mode Water continues to be a major component of the Current's transport which is now basically to the east.

[3] Note that the general definition (Iselin, 1936) of the North Atlantic Current is used here. Clarke *et al.* (1980) use a narrower definition, restricting the North Atlantic Current to the northward flow which originates from branching of the combined Gulf Stream and Slope Water Current west of the Tail of the Grand Banks. The remainder of the northward flow, in their terminology, is attributed to the westward limb of a "Newfoundland Basin eddy." They indicate that this limb contributes about 40% of the total northward flow. In this terminology, both the North Atlantic Current and the Newfoundland Basin eddy transport Mode Water northward.

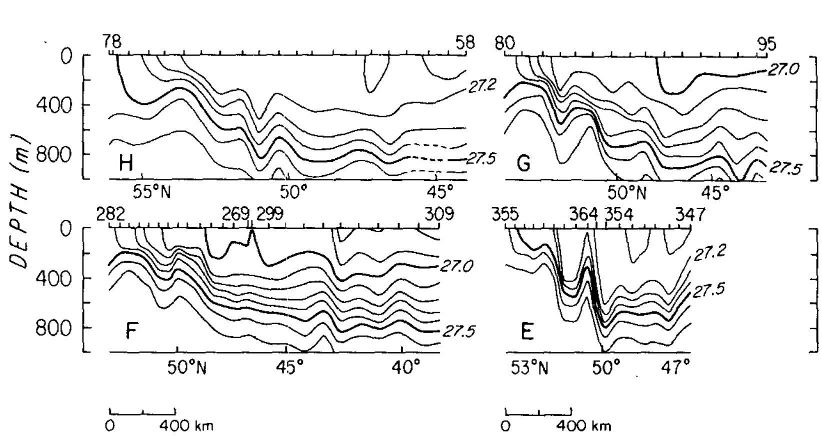

FIG. 9. Potential density for sections E–H (Fig. 5b and Appendix Table A2). These sections are oriented meridionally and illustrate the eastward transport of the North Atlantic Current after it separates from the western boundary near section E. Overlying the north to south sloping pycnocline that indicates this eastward flow are thick layers at densities in the range 27.0 mg cm^{-3} $\leqslant \sigma_\theta \leqslant$ 27.2 mg cm^{-3}. These pycnostads are thus indicated as moving eastward with the Current.

The Mode Water in the Current is not just simply advected by the Current. It also changes characteristics, becoming denser, colder and fresher downstream. The presence of vertical homogeneity implies vertical convection, but the air-sea fluxes driving such convection need not be the only fluxes producing the downstream trends. Cross-current eddy fluxes of heat and salt could also contribute, with local convection being a seasonal homogenizing agent.

Changes also occur as a function of time. In Fig. 10, section F′ was occupied in April 1964, about 2½ months after section F in Fig. 8, but at the same location. The deep mixed layer overlying the Current has deepened and increased in density. Similarly, section C′ (Fig. 10) was occupied in June 1964, about four months after section C in Fig. 8, but at nearly the same location. The pycnostad in C′ is deeper and denser than that adjacent to the Current in section C.[4]

Part of the Mode Water transported by the northward and eastward flow of the North Atlantic Current in sections C–H recirculates in the anticyclonic regime east and south of the Current. The southward recirculation is illustrated in Fig. 11 by a set of six density sections. The southern five are zonal lines, but the northernmost one, "48°N," is slightly oblique (Fig. 5b). It runs from 45°N in the west to 50°N in the east, and lies within the deep convection region all the way from Newfoundland to Ireland.

All six sections show an increase of isopycnal spacing in the density range σ_θ = 26.8–27.25 mg cm^{-3}. The Mode Water influence at 48°N is strong and clear in this April–May section. In the west is the northward flowing Current, transporting light Mode Water, with a compact southward recirculation of slightly heavier Mode Water offshore. Over the eastern part of the section there is a broad region where isopycnals slope upward to the east, indicating weak southward shear at the level of the Mode Water pycnostad. Saunders (1982) indicates that a deep reference level (~4000 db) gives total, depth-integrated transports consistent with those predicted by the distribution of wind-stress curl (Leetmaa and Bunker, 1978). Using the deep reference level south of 50°N, the ocean is conveniently broken into three layers on the basis of geostrophic shear. The shear above 850 m, the presence of the Mode Water pycnostad at densities above σ_θ = 27.3 mg cm^{-3}, and the reference level combine to indicate a southward flow of Mode Water in the upper 850 m of the eastern North At-

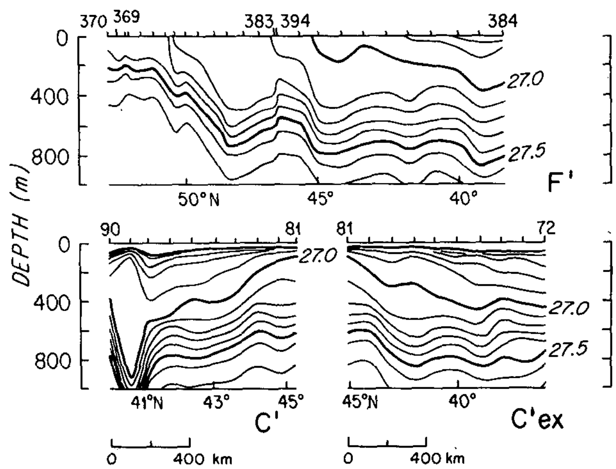

FIG. 10. Potential density for section C′, C′$_{ex}$ and F′ (Fig. 5b and Appendix Table A2). C′ and F′ are later reoccupations of sections C and F in Fig. 9. The pycnostads at 26.8 mg cm^{-3} $\leqslant \sigma_\theta \leqslant$ 27.2 mg cm^{-3} are the Subpolar Mode Water. C′$_{ex}$ illustrates that pycnostads flow eastward across 35°W in late spring as far south as 38°W.

[4] Station 89 contains an extremely pronounced pycnostad overlying an anomalously deep pycnocline. This could be interpreted as an extreme example of anticyclogenesis (Worthington, 1972). It may be evidence of Mode Water formation as mesoscale convection "chimneys" (Killworth, 1979).

lantic. Fig. 12a shows Saunders' (1982) estimate of the pattern of the transport for the 0–850 m layer.

The pycnostads in the eastern-basin parts of the sections in Fig. 11 fall in the density range σ_θ = 27.0–27.25 mg cm^{-3}. This corresponds to the density of deep convection in sections E, F, G and H in Fig. 9 (and F′ in Fig. 10). Some of this eastward transport of Mode Water is thus indicated as recirculating southward basinwide—an anticyclonic recirculation. The pycnostads are quite pronounced at the 48, 46, 43 and 40°N sections. At 36°N the pycnostads have weakened considerably, and by 32°N they are barely discernable given the bottle spacing (~100 m) of the data. This limitation to 10 or 15° of latitude for the pycnostad influence from a deep convection source at higher latitudes is similar to that of the Eighteen Degree Water in the western subtropical Atlantic.

The southward circulation across 43, 40, 36 and 32°N in the Saunders interpretation is fed by eastward and southeastward flow across the mid-ocean ridge in the neighborhood of sections F and F′ (Figs. 9 and 10). This is reflected by the continuation of the north-to-south descent of the pycnocline from the main axis of the Current all the way south to the Azores. The 35°W continuation of section C′ (Fig. 10) also shows a north-to-south descent of the pycnocline indicating eastward flow.

The notion of southward flow in the upper water column of the eastern North Atlantic Ocean is, of course, not new (Sverdrup *et al.*, 1942; Iselin, 1936; Helland-Hansen and Nansen 1926). The present direct evidence for this flow is the advection of pycnostads from a deep convective source region at higher latitudes. The pycnostads in Fig. 11 seem to have gone unnoticed, probably because of the complicating influence of the Mediterranean water immediately underneath the pycnostads. In Fig. 11 the light shading indicates the depth range for each station where salinity increases with depth: the transition layer to the Mediterranean water core. The temperature and salinity sections of Fuglister (1960) for these same stations show thermostads and halostads spanning the 300–1500 m depth range. These encompass both the Mode Water and the Mediterranean Water. The convective origin of the upper parts of the thermostad and halostad layers becomes evident only by the relative vertical homogeneity in density.

Combining the Saunders (1982) interpretation of the upper and middle layer circulations (Figs. 12a and 12b) with the density sections of Fig. 11 gives an image of an active southward flow of pycnostads in the upper layer above a relatively motionless Mediterranean Water. In Fig. 11, the shaded layers vertically separate the Subpolar Mode Water pycnostads from the core of the Mediterranean Water. This transition layer is centered near 850 m. The Mode Water pycnostads thus lie in the relatively active 0–850 m layer (Fig. 12a), while the main Mediterranean Water influence is in the comparatively quiescent 850–1200 m layer (Fig. 12b). It is this southward advection of

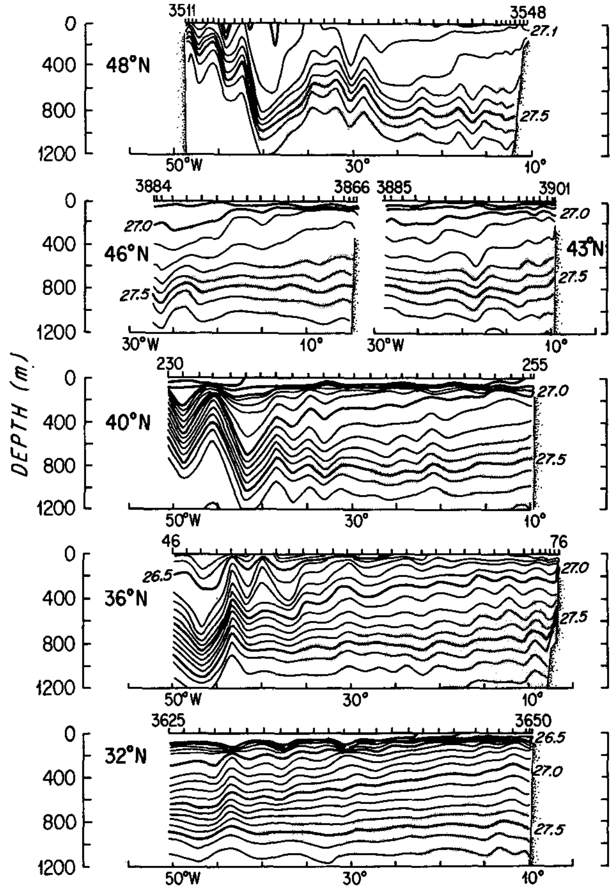

FIG. 11. Potential density for sections at the indicated latitudes: 32, 36, 40, 43, 46°N and "48°N"—the last an oblique section from about 45°N in west to 50°N in east (Fig. 5 and Appendix Table A2). East of 30°W the pycnostad is in the density range 27.0 mg cm^{-3} $\leqslant \sigma_\theta \leqslant$ 27.3 mg cm^{-3}. In the northern four sections the pycnostad is strong, at 36°N it is noticably weaker, and at 32°N it is barely discernible as about a 20% excess thickness of isopycnal spacing. The shaded area indicates where salinity increases with depth—the transition layer between the pycnostad, of convective origin, and the underlying Mediterranean Water. West of 30°W the 48°N and 40°N sections show the North Atlantic Current and a local recirculation of Mode Water. The Gulf Stream transports Eighteen Degree Water at $\sigma_\theta \approx$ 26.5 mg cm^{-3} southward through the western ends of the 36 and 32 N sections.

pycnostads that provides the high-volume ridge in the 10–15°C range in the North African Basin (Fig. 3).

Within the ocean-wide anticyclonic recirculation there appears to be a western-intensified compact component. Crossing the North Atlantic Current from left to right (looking downstream), the pycnocline steeply descends, reaching a maximum depth at the right edge (by definition) of the Current. Beyond that edge the pycnocline slope reverses. In the west that region of reversed slope is sometimes quite steep, for example in sections C and D (Fig. 8), E (Fig. 9), C′ (Fig. 10), and 48°N and 40°N (Fig. 11). This is a compact recirculation of part of the North Atlantic Current. The southern and eastern extent of this compact feature may be time-dependent. For example, section F (Fig. 8) does not show the slope

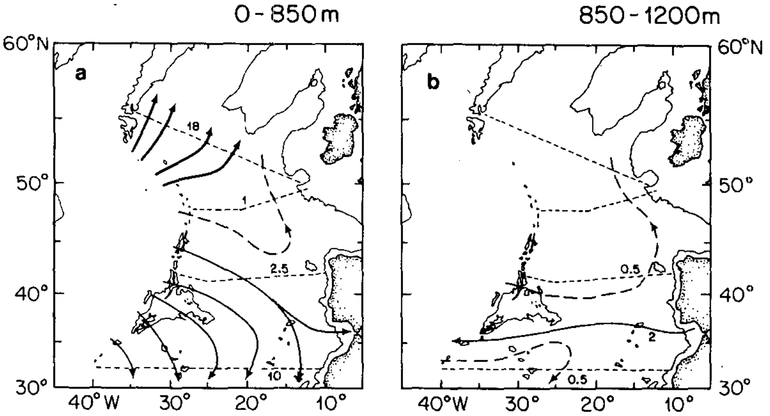

FIG. 12. The Saunders (1982) interpretation of the transport (10^6 m^3 s^{-1}) pattern in the eastern North Atlantic, for the data sections shown by dashed contours. Three of the four sections used are illustrated in Figs. 11 and 13 of the present paper: sections 32°N, 48°N, and I. (a) 0–850 m, (b) 850–1200 m.

reversal, while F′ (Fig. 10), 2½ months later, does. Section B (Fig. 8) contains a southeastward flowing branch of the Gulf Stream on its western end, but no sign of the northward flowing Current nor a compact recirculation therefrom, nor an overlying Mode Water. On the other hand, the section on 36°N in Fig. 11, overlapping the location of B, shows some indication of the Current/compact recirculation geometry, centered at 40°W, with a Mode Water of σ_θ = 26.85 mg cm^{-3}.

The anticyclonic recirculation just described agrees in general terms with that implied by the discussion of Clarke *et al.* (1980) for the region east of Newfoundland. It disagrees with that proposed by Worthington (1976) in that it is ocean-wide in extent. Clarke *et al.* (1980) estimate that about 40% of the northward transport recirculates in a compact "Newfoundland Basin Eddy" located west of 40°W. The implication then is that the remaining 60% flows eastward to recirculate anticyclonically east of 40°W, or perhaps with part turning north into the cyclonic regime. They do not explicitly discuss that point. Their compact "eddy" was estimated at 18–25 × 10^6 m^3 s^{-1} (upper 2000 m). The remaining 44–66 × 10^6 m^3 s^{-1} passes east past 40°N. Saunders (1982) estimates an eastward transport crossing the mid-ocean ridge of about 29 × 10^6 m^3 s^{-1} in the upper 850 m. The pycnostad distributions in Fig. 11 indicate the anticyclonic recirculation extends west of the ridge. Hence, southward recirculation between 40°W and the mid-ocean ridge could account for the difference between 29 and 44–66 × 10^6 m^3 s^{-1}. A consistent picture thus emerges: of the total northward Current transport off Newfoundland, part recirculates west of the mid-ocean ridge, while the remainder passes eastward into the eastern basin, to participate either in the anticyclonic recirculation there, or to branch northward.

Worthington (1976) interpreted the transport of the Current as almost totally recirculating anticyclonically west of 30°W. Only relatively small amounts of water were indicated as flowing east across the mid-ocean ridge, and most of that turns north rather than recirculating southward in the eastern basin. Various arguments were used in Worthington's (1976) monograph to support this lack of anticyclonic recirculation east of 30°W, and we will comment here on two of those: the oxygen level in the main pycnocline and the Mediterranean Water.

Worthington's concern with the Mediterranean Water influence was to avoid a major volume transport passing through the tongue. The pycnostads illustrated in Fig. 11 do not pass through the tongue, they pass over the tongue, as described above. The pycnostad salinity anomalies relative to western Atlantic standard salinities are much smaller than those of the Mediterranean Water, only about +0.02 to +0.10‰ (Fig. 3). These levels of salinity anomaly *are* found in the western Atlantic subtropical gyre, for example at 55°W (McCartney *et al.*, 1980). It is therefore possible to have ocean-wide recirculation without doing violence to basic water-mass distributions.

A second concern of Worthington involved the oxygen levels in the thermocline. East of Newfoundland, in the density range $\sigma_\theta \leqslant 27.5$ mg cm^{-3}, oxygen levels are considerably higher than in the western subtropical Atlantic. This may reflect cross-Current mixing as suggested by Clarke *et al.* (1980), and measurements by Georgi and Schmitt (1982) indicate the necessary finestructure intensity to support such mixing. An additional contributing factor is the recirculation of the pycnostads. At $\sigma_\theta \leqslant 27.25$ mg cm^{-3}, convectively formed pycnostads recirculate anticyclonically. Convection imparts high oxygen content to the water, effectively the saturation value. The pycnostads are thereby high-oxygen-content water masses. Many people (Mann, 1967, for example) have referred to the eastern subtropical Atlantic as being closer to the outcropping source for high oxygen. In general these higher levels are anticipated in the east.

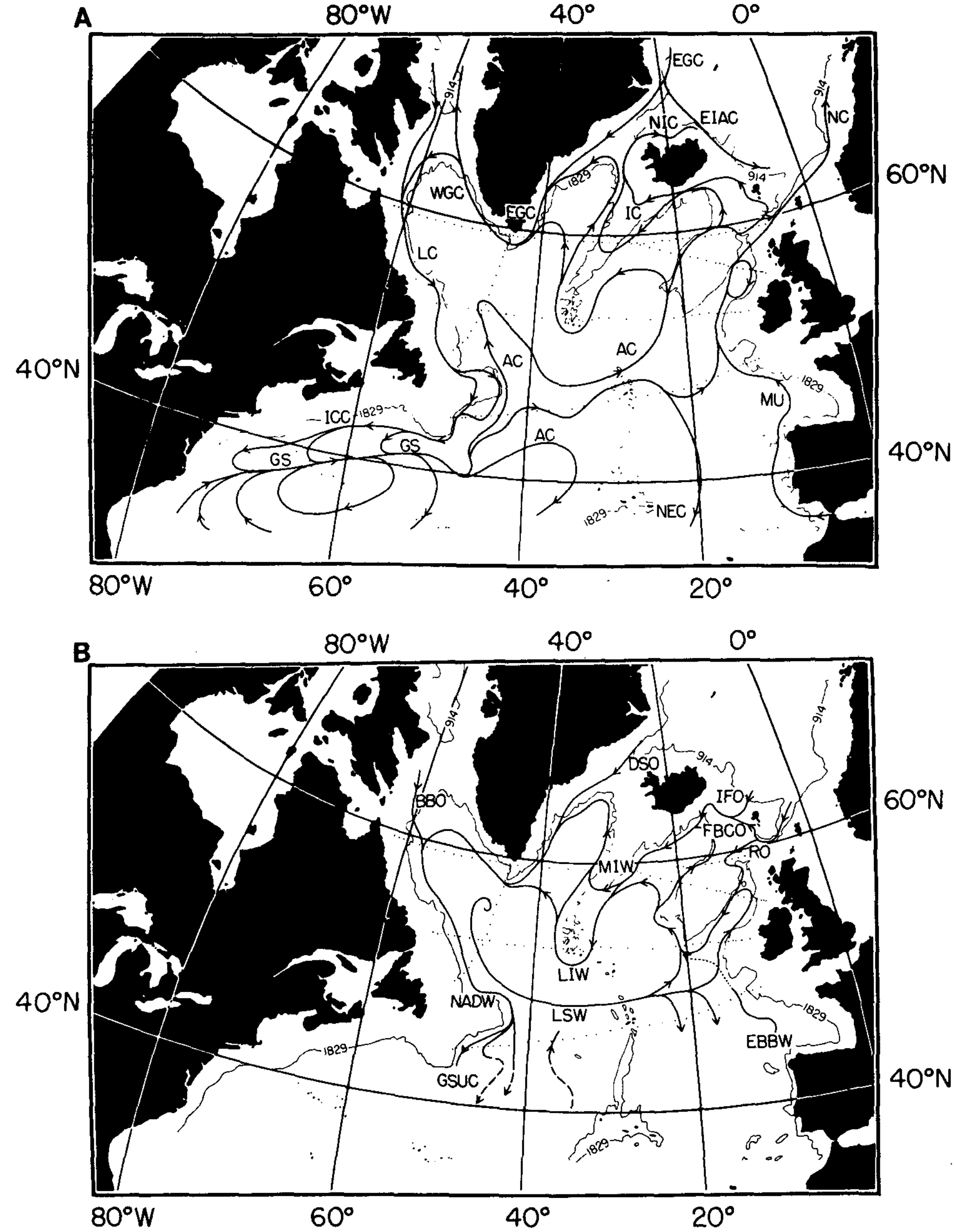

FIG. 13. The Ivers (1975) interpretation of the circulation pattern for the northern North Atlantic. Panel A is the upper water, Panel B the deep water. The isobath values, in meters, represent conversions from round numbers of fathoms.

An additional factor is that the compact recirculation feeds the oxygenated pycnostads directly back into the North Atlantic Current, giving higher oxygen levels (and pycnostad strengths) east of Newfoundland compared to those south of Newfoundland.

In summary, the material discussed in this subsection shows northward flow of Mode Water in the North Atlantic Current off Newfoundland, turning eastward near 52°N and extending eastward across the Atlantic into the eastern basin. Along this advection path the Mode Water density increases and the salinity and temperature decrease. Part of the Current feeds anticyclonic recirculation to the south and east of the Current. This recirculation may include a compact gyre or eddy west of the mid-Atlantic ridge and less intense southward flow east of the ridge. The remainder of the eastern flowing Current crossing the mid-ocean ridge turns north, as described next.

b. *Transport of Mode Water in the subpolar cyclonic circulation*

The separated North Atlantic Current flows eastward across the North Atlantic, crosses the mid-ocean

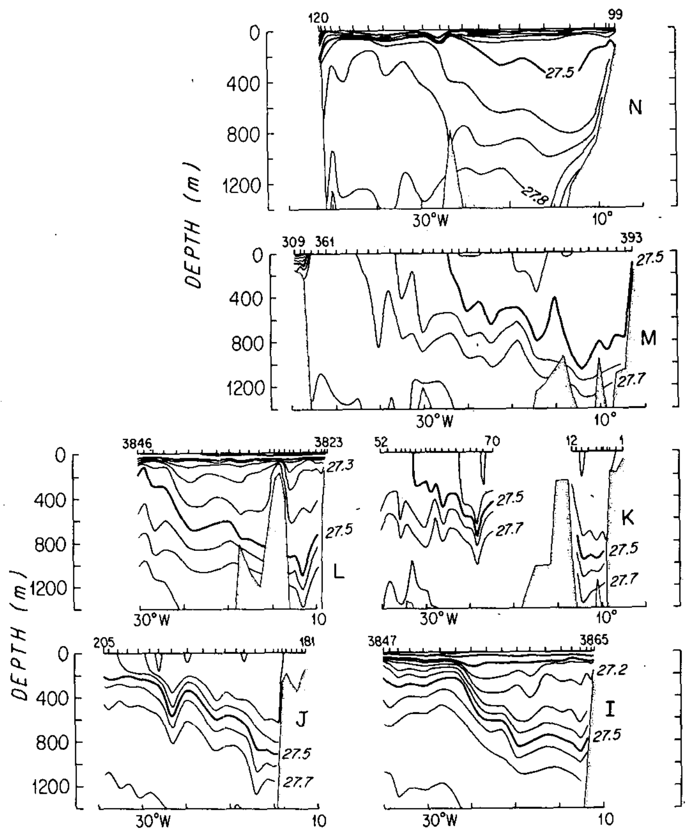

FIG. 14. Potential density for Sections I–N in the eastern subpolar North Atlantic (Fig. 5b and Appendix Table A2). East of 25°W the west-to-east descent of the pycnocline is indicative of northward transport. Overlying this sloped pycnocline are the Subpolar Mode Water pycnostads. These pycnostads increase in density from south to north, beginning in sections I and J at $\sigma_\theta \leqslant 27.3$ mg cm^{-3}, rising to $\sigma_\theta \leqslant 27.4$ mg cm^{-3} at the latitude of the Rockall Channel (sections K and L; see also Fig. 4), reaching $\sigma_\theta \leqslant 27.5$ mg cm^{-3} just south of the Faroe-Shetland Channel and the Iceland Faroe Ridge (section M) and $\sigma_\theta \approx 27.5$ mg cm^{-3} at 62°N (section N).

ridge and divides into northward and southward flows in the eastern basin. This bifurcation is an old concept in the description of the general circulation of the North Atlantic. Sverdrup *et al.* (1942, Fig. 187) describe such a pattern, as did Iselin (1936, Fig. 48) and Helland-Hansen and Nansen (1926, Fig. 15). In the preceding subsection, the participation of Subpolar Mode Water pycnostads in the southward turning branch was described. In the present subsection, the northward turning branch is discussed, and the continuation of the progression of denser pycnostads in the Subpolar North Atlantic is described.

Ivers (1975) thoroughly analyzed the northern North Atlantic circulation, and his interpretation of the shallow and deep circulation is reproduced here as Fig. 13. He used geostrophic shear calculations for individual station pairs and a qualitative assessment of likely flow direction as a function of depth, based on water-mass distributions. He found a tendency for columnar flow: flow in the same direction from top-to-bottom for a given station pair. This is seen in Fig. 13 in the parallel advection paths in the upper and lower water column north of 50°N. North of 50°N he deduced a cyclonic subpolar gyre, fed by the North Atlantic Current (NAC).

The northward flow past 50°N in the eastern Atlantic envelops the Rockall Bank. Sections I–N in Fig. 14 illustrate this general northward flow. The pycnocline slopes downward from near 25 or 30°W to the continental slope, indicating northward flow relative to a deeper reference level. Section I, the "53°N" section Saunders (1982) used, shows a pycnostad at $\sigma_\theta \approx 27.25$ mg cm^{-3} overlying this sloped pycnocline and hence embedded in the northward flow. Section J shows a similar configuration, with no seasonal pycnocline capping the pycnostad in this winter section (recall the seasonal contrast illustrated in Fig. 4). The pycnostad layer thus continues to contribute a sizable fraction of the transport of this branch of the Current.

Following this northward flow through the Rockall Bank region, sections K–M show the pycnostad density gradually increasing, reaching $\sigma_\theta \approx 27.45$ mg cm^{-3} at section M. In the east, section M terminates immediately south of the Faroe Bank Channel, and a branch from the northward flow penetrates the Channel to provide the warm core of the Norwegian Current (NC in Fig. 13). A section across this Current immediately north of the Faroe-Shetland sill is shown in Fig. 15. The section (NC) was occupied about three weeks after section M, and shows a pycnostad at $\sigma_\theta \approx 27.5$ mg cm^{-3} flowing northward on the inshore side of the Norwegian Current.

Immediately south of the Iceland and the Iceland-Scotland Ridge (east of 25°W), section N in Fig. 14 shows the pycnostad density at about $\sigma_\theta \approx 27.5$ mg cm^{-3}. In this region the northward flow turns west, becoming the Irminger Current (Fig. 13, IC). This Current is deflected southwest by the Reykjanes Ridge. This deflection shows as a reversal in pycnocline slope in sections M and N over the east flank of the Ridge. In this southwestward flow the pycnostad density is even higher, falling near $\sigma_\theta = 27.55$ mg cm^{-3} in these two sections.

The Ivers (1975) upper-water circulation pattern shown in Fig. 13a shows this meander extending all the way southwest of 35°W on what is section J in Fig. 14. The flow then turns back northward along the west flank of the Ridge. The Reykjanes Ridge meander is more completely discussed by Langseth and Boyer (1972).

The $\sigma_\theta \approx 27.55$ mg cm^{-3} pycnostad that accompanies the southwestward flow can be seen in the neighborhood of the ridge crest in sections L and K. Section J does not show any pycnostad influence at 35°W, so it appears that the flow of pycnostads crosses the ridge north of 53°N. The northwest end of section H in Fig. 9 near this same location also shows a clear pycnostad of $\sigma_\theta \approx 27.55$ mg cm^{-3}. This section is helpful for it shows this pycnostad to be

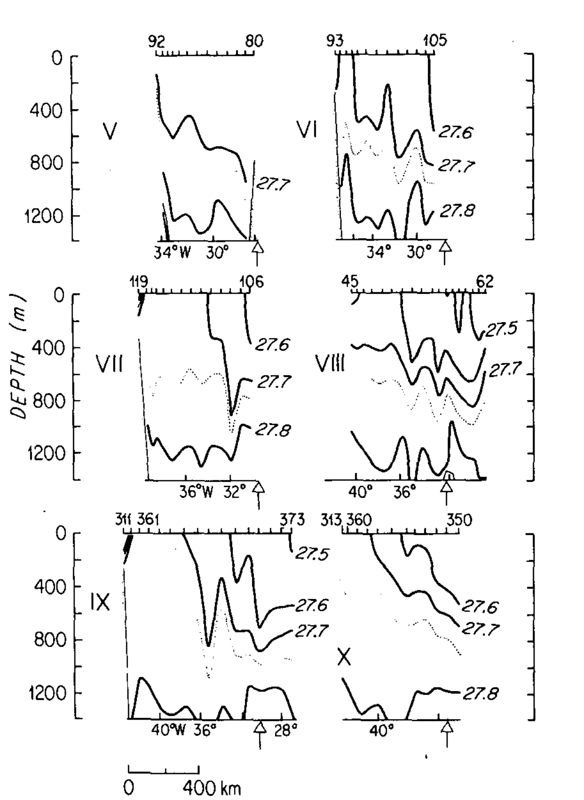

FIG. 16. Potential density for sections V–X between the east coast of Greenland and the Reykjanes Ridge (Fig. 5b and Appendix Table A2). The arrow is the approximate location of the Ridge crest. On the west flank of the Ridge a northward transport occurs, with the Mode Water pycnostad becoming denser from south to north—from $\sigma_\theta \leq 27.6$ gm cm^{-3} to $\sigma_\theta \leq 27.7$ mg cm^{-3}. This northward flow turns cyclonically westward near section IV, Fig. 17, then southward through the west half of sections V–X, continuing to gain in density, and reaching $\sigma_\theta \approx 27.75$ mg cm^{-3} at section X, and the south tip of Greenland.

well north of the eastward flow of the North Atlantic Current (south of 53°N). It appears from these sections that maximum southwest penetration of the flow of pycnostads before turning back northward is to the general neighborhood of 35°W, 55°N.

The northward flow on the west flank of the Reykjanes Ridge carries still denser Mode Water pycnostads. Sections V–X (Fig. 16), spanning the Irminger Sea from the east coast of Greenland to the Reykjanes Ridge, show the pycnostads rising from $\sigma_\theta \approx 27.55$ mg cm^{-3} in the south to $\sigma_\theta \approx 27.65$ mg cm^{-3} in the north.

South of the Denmark Strait, the Irminger Current branches, with a northward branch carrying Mode Water pycnostads through the Strait to provide a warm core to the North Icelandic Irminger Current (Stefánsson, 1962; also NIC in Fig. 13). Sections I–IV in Fig. 17 illustrate the penetration of pycnostads through the Strait, above the southward flowing Denmark Strait overflow water.

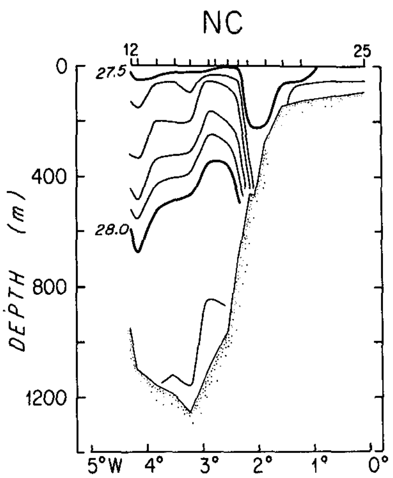

FIG. 15. Potential density for section NC just north of the Faroe-Shetland Ridge (Fig. 5b and Appendix Table A2). Section occupied three weeks after the eastern end of section M in Fig. 13. The Norwegian Current lies between stations 18 and 24, and contains a pycnostad at $\sigma_\theta \approx 27.5$ mg cm^{-3}. This is Subpolar Mode Water from the open North Atlantic that flows northwards through the Faroe-Shetland Channel.

The remaining flow of the Irminger Current turns cyclonically around the Irminger Sea and becomes part of the southward flow of the East Greenland

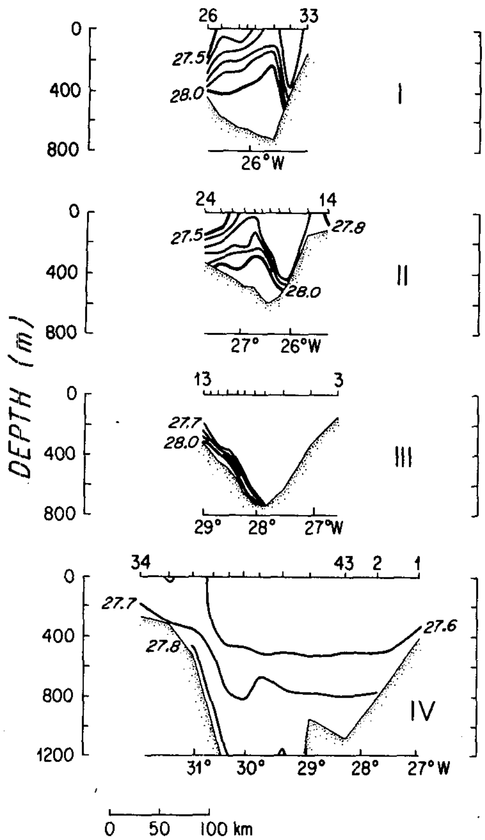

FIG. 17. Potential density for sections I–IV, near the Denmark Strait (Fig. 5b and Appendix Table A2). At section IV the main flow of Mode Water is east to west, but a northward branch flows from the neighborhood of section IV through sections I–III, providing a warm core to the North Icelandic Irminger Current. The core potential density has risen to $\sigma_\theta \approx 27.7$ mg cm^{-3} at the northernmost section I.

Current (EGC in Fig. 13). The pycnostads forming a warm core in the offshore part of this current continue to increase in density in the flow direction, rising from $\sigma_\theta \approx 27.65$ mg cm^{-3} at section V to $\sigma_\theta \approx 27.75$ mg cm^{-3} at the southern end of Greenland, the northwest end of section X (Fig. 16). The pycnostad temperature drops below 5°C along this advection path.

At the southern tip of Greenland, the westward flow of Mode Water pycnostads provides a warm core to the West Greenland Current. After penetrating partway into the Davis Strait, the core turns westward and then southward as part of the Labrador Current. The pycnostad lying in the offshore parts of the West Greenland and Labrador Currents is readily seen in the oceanographic atlases of this region (Worthington and Wright, 1970; Grant, 1968) as a warm, salty core beginning at the southern tip of Greenland warmer than 4°C but cooling below 4°C in the Labrador Current. The progression of denser pycnostads ends with this cyclonic current system in the Labrador Sea, with maximum pycnostad densities of $\sigma_\theta \approx 27.78$ mg cm^{-3}.

The properties of the densest Subpolar Mode Water pycnostads are those of the Labrador Sea Water. The example in Fig. 7a has potential temperature $\theta \approx 3.4$°C and salinity $S \approx 34.88$‰. The Labrador Sea Water core characteristics change from year to year. A companion study by Talley and McCartney (1982) discusses the interannual variability of the Labrador Sea Water, and its influence at mid-depth throughout the North Atlantic. Because the Labrador Sea Water is the last, densest variety of Subpolar Mode Water, its properties depend on the entire history of the Subpolar Mode Water as it is advected cyclonically around the subpolar North Atlantic, including the path history of air–sea and lateral heat exchange.

The existence of a cyclonic subpolar circulation is presumably a consequence of the wind-stress distribution of the North Atlantic. The wind-stress curl according to Leetmaa and Bunker (1978) is zero near 55°N off Ireland, i.e., in the region of the northward branch of the North Atlantic Current. To the north, the wind-stress curl is positive, resulting in northward Sverdrup flow and hence cyclonic circulation; to the south, the wind drives an anticyclonic gyre. This basic pattern agrees with the pycnostad distributions and circulations we have described in this paper. An estimate of the transport of the wind-driven cyclonic circulation is given by Leetmaa and Bunker (1978) as 40×10^6 m^3 s^{-1}, based on their calculation of wind-stress curl. This is the same as Ivers' (1975) transport estimate, which was obtained in a completely independent manner.

In summary, there is a northward flow of Subpolar Mode Water off Ireland which originates from a branch of the North Atlantic Current and supplies warm water to the subpolar North Atlantic. Particular pycnostads enter the Norwegian Sea through the Faroe-Shetland Channel and the Denmark Strait and provide the warm core of the Norwegian Current and North Icelandic Irminger Current, respectively. The remaining warm water flows from east to west across the Subpolar North Atlantic, with the pycnostads progressively cooling from warmer than 7°C southeast of Iceland to cooler than 4°C in the Labrador Sea. The main axis of this advection is the sequence of northern boundary currents: the Irminger, East Greenland, West Greenland and Labrador Currents. The Subpolar Mode Water is the warm core of each of these currents. In a companion study (McCartney and Talley, 1982) we discuss estimates of the warm water to cold water conversion rate for the region north of 50°N.

Acknowledgments. This work was supported by the Office of Naval Research under Contracts N00014-74-C-0262, NR 083-004, and N00014-79-C-0071, NR 083-400 and by the International Decade of Ocean Exploration, Office of the National Science Foundation under Grants OCE76-00390 and OCE78-22885. The authors wish to express their appreciation to Val Worthington for many helpful discussions, to

Bruce Warren for many useful suggestions as to presentation, to Mary Raymer for her assistance in the analysis and preparation of the materials for this study, and to Doris Haight and Karin Bohr for their diligence in preparing the manuscript. This is Woods Hole Oceanographic Institution Contribution No. 4614.

APPENDIX

Tables A1 and A2 summarize the ships, station numbers and dates of the data used in, respectively, the late-winter charts and the various sections. Included are references which contain various property sections and/or charts using the same data.

TABLE A1.*

Symbol	Ship	Date	Stations	References
■	*Anton Dohrn*	Mar 1958	2238–2300	Dietrich (1969)
▽	*Explorer*	Mar 1958	52–100	—
		Apr 1957	12–41	—
◇	*Helland Hansen*	Mar 1958	1–71	—
▲	*Ekvator*	Mar–Apr 1958	51–69	Dietrich (1969)
△	*Lomonosov*	Mar 1958	52–95	Dietrich (1969)
⍋	*Gauss*	Apr 1958	44–55	—
○	*Erika Dan*	Jan–Apr 1962	181–393	Worthington-Wright (1970)
◇	*Dalrymple*	Mar 1964	6–12	Ellett-Martin (1973)
×	*Atlantis II*	Jan–Apr 1964	284–418	Worthington (1976)
+	*Hudson*	Jan–Mar 1967	1–119	Grant (1968)
	Gulf Stream 60	Apr 1960		Fuglister (1963)
□	*Chain* 12		136–202	
	Crawford 40		809–870	
	Atlantis 255		5873–5926	
	Evergreen		7294–7309	

* See Fig. 5a.

TABLE A2.*

Section	Ship	Date	Stations	References
A	*Evergreen*	Apr 1960	7300–7293	Fuglister (1963)
B	*Atlantis II*	Feb 1964	310–325	Worthington (1976)
C	*Atlantis II*	Mar 1964	326–346	Worthington (1976)
C′	*Baffin*	Jul 1964	81–90	Mann (1967)
C′ex	*Baffin*	Jul 1964	72–81	Mann (1967)
D	*Atlantis II*	Apr 1964	394–410	Worthington (1976)
E	*Atlantis II*	Mar–Apr 1964	347–364	Worthington (1976)
F	*Atlantis II*	Jan–Feb 1964	269–282	Worthington (1976)
			299–309	
F′	*Atlantis II*		369–394	Worthington (1976)
G	*Lomonosov*	Mar 1958	80–95	Dietrich (1969)
H	*Lomonosov*	Mar 1958	78–58	Dietrich (1969)
I	*Discovery*	Aug 1958	3847–3865	Fuglister (1960)
J	*Erika Dan*	Jan–Feb 1962	181–205	Worthington-Wright (1970)
K	*Dalrymple*	Mar 1964	1–10, 12	Ellett-Martin (1973)
	Hudson	Feb 1967	52–70	Mann (1969)
L	*Discovery*	Aug 1958	3823–3846	Fuglister (1960)
M	*Erika Dan*	Mar–Apr 1962	309–313	Worthington-Wright (1970)
			361–393	
N	*Dana*	Jul 1958	99–120	
NC	*Helland Hansen*	May 1962	12–25	
I–VII	*Hudson*	Jan–Apr 1967	1–119	Mann (1969)
IX	*Erika Dan*	Mar–Apr 1962	311–313	Worthington-Wright (1970)
			361–373	
X	*Erika Dan*	Mar–Apr 1962	313	Worthington-Wright (1970)
			350–360	
32°N	*Discovery*	Nov–Dec 1957	3625–3650	Fuglister (1960)
36°N	*Chain*	Apr–May 1959	46–76	Fuglister (1960)
40°N	*Crawford*	Oct 1957	233–255	Fuglister (1960)
43°N	*Discovery*	Sep 1958	3885–3901	Fuglister (1960)
46°N	*Discovery*	Sep 1958	3866–3884	Fuglister (1960)
48°N	*Discovery*	Apr 1957	3511–3548	Fuglister (1960)

* See Fig. 5b.

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