Thermohaline Circulation in the Arctic Mediterranean Seas

K. AAGAARD
School of Oceanography, University of Washington, Seattle

J. H. SWIFT
Scripps Institution of Oceanography, La Jolla, California

E. C. CARMACK
Department of the Environment, West Vancouver, British Columbia

The renewal of the deep North Atlantic by the various overflows of the Greenland-Scotland ridges is only one manifestation of the convective and mixing processes which occur in the various basins and shelf areas to the north: the Arctic Ocean and the Greenland, Iceland, and Norwegian seas, collectively called the Arctic Mediterranean. The traditional site of deep ventilation for these basins is the Greenland Sea, but a growing body of evidence also points to the Arctic Ocean as a major source of deep water. This deep water is relatively warm and saline, and it appears to be a mixture of dense, brine-enriched shelf water with intermediate strata in the Arctic Ocean. The deep water exits the Arctic Ocean along the Greenland slope to mix with the Greenland Sea deep water. Conversely, very cold low-salinity deep water from the Greenland Sea enters the Arctic Ocean west of Spitsbergen. Within the Arctic Ocean, the Lomonosov Ridge excludes the Greenland Sea deep water from the Canadian Basin, leaving the latter warm, saline, and rich in silica. In general, the entire deep-water sphere of the Arctic Mediterranean is constrained by the Greenland-Scotland ridges to circulate internally. Therefore it is certain of the intermediate waters formed in the Greenland and Iceland seas which ventilate the North Atlantic. These waters have a very short residence time in their formation areas and are therefore able to rapidly transmit surface-induced signals into the deep North Atlantic.

Introduction

The primary northern hemisphere source of deep ventilation for the World Ocean lies north of the Greenland-Scotland ridge system, over which dense water spills into the deep North Atlantic (cf. Mantyla and Reid [1983] for a comprehensive discussion). The seas to the north of these ridges consist of a series of interconnected basins, each with its own distinctive characteristics and contributions to the thermohaline circulation. In keeping with earlier nomenclature, we denote these basins collectively as the Arctic Mediterranean (AM) [cf. Sverdrup et al., 1942, p. 15], and in our discussion of their ventilation, we shall pay particular attention to the role of the Arctic Ocean. Far from being simply a passive recipient of ventilated water from the south, we shall show that the Arctic Ocean is itself an important source of dense water, a portion of which is exported southward through Fram Strait. This is a role in the thermohaline circulation distinctly different from the estuarine one which in the past has received the principal attention [e.g., Stigebrandt, 1981].

Figure 1 shows that the Arctic Ocean constitutes by far the largest portion of the AM. There are two major basins, the Canadian and Eurasian, bordered by extensive shelf seas. South of 2600-m-deep Fram Strait, the basin complex west of the mid-ocean ridge is defined as the Greenland Sea and that to the east is the Norwegian Sea. The latter leads into Fram Strait through a long trough extending northward. The area between Iceland, Greenland, and the island of Jan Mayen has its own distinctive circulation and hydrography, and it is usually referred to as the Iceland Sea. The ridges from Greenland to Scotland which confine the AM at its southern end are relatively shallow, being about 600-800 m at their deepest.

The hypsography of the AM (Figure 2) shows the total volume to be $17 \times 10^6 \text{ km}^3$, or about 1.3% of the volume of the World Ocean. The largest components are the two major Arctic Ocean basins, which make up 75% of this volume; the Canadian Basin alone accounts for 43% of the total volume of the AM. Figure 2 also makes clear the large proportion of the Arctic Ocean which is continental shelf. The Greenland and Norwegian seas together account for only 22% of the total volume, and they do not have unusually large shelves. The Iceland Sea contains only about 2% of the volume, but it is of major importance in ventilating the North Atlantic [Swift et al., 1980].

Figure 2 also gives the mean depth of selected isopycnals within each basin of the AM (cf. the appendix for discussion of notation and units used in this paper and a comparison at different pressures of the isopycnals of Figure 2). The surface $\sigma_0 = 27.9$ separates the upper water masses, including those of the pycnocline, from the intermediate ones in the next density range. Although it is these uppermost waters which have been most heavily studied, they constitute only 17% of the total volume of the AM.

The intermediate waters have densities as great as $\sigma_1 = 32.785$, which value is found at the sea surface in the cyclonic gyres of the Greenland and Iceland seas during winter, while waters denser than this outcrop much more rarely. The division at this density value was in large part chosen because no denser water is directly exported to the North Atlantic. The mean lower boundary of these intermediate overflow waters is found at very shallow levels in both the Greenland and Iceland seas. The intermediate waters comprise 29% of the total volume, so that 46% of the total volume of the AM is potentially in communication with the remainder of the World.
Ocean, while 54% is effectively isolated. Because the traditional deep-water boundary of 0°C lies mostly above the $\sigma = 32.785$ surface, the estimate that 54% of the total volume is made up of deep water is by traditional standards an underestimate.

In Figure 2 we have distinguished between deep waters in the density range $\sigma_1 = 32.785$ to $\sigma_2 = 37.457$ and those denser than $\sigma_2 = 37.457$. The latter owe their high density to the relatively low temperatures, near or below $-1^\circ$C. This is particularly important at elevated pressures, because of the tem-
perature dependence of the compressibility. In the Arctic Ocean the Lomonosov Ridge rises above the 37.457 \( \sigma_t \)-surface, and the densest waters in the Eurasian Basin are therefore excluded from the Canadian Basin. Furthermore, deep-water formation in the Canadian Basin apparently does not yield a product cold enough to raise the density to the highest values found in the other major basins. The seeming absence of the denser forms of deep water from the Iceland Sea (Figure 2) is due to the very small amounts of such water present in this shallow basin.

The effectiveness of surface-driven convection in ventilating the deep ocean depends critically on the prevailing stratification and associated ice conditions. The Arctic Ocean is in fact so stably stratified by low-salinity surface water that winter cooling does not drive convection over the deep basins below about 50 m. Conversely, ice formation is suppressed when saline water is supplied to a region of intense cooling, destabilizing the water column. This appears to be the case in the Greenland Sea, which not only receives cold, low-salinity surface water from the Arctic Ocean via the southward flowing East Greenland Current, but also is fed by warm and saline water from the south via the Norwegian Atlantic Current. The surface water entering the Norwegian Sea from the Atlantic is amply saline to form deep water instead of freezing, but it is initially much too warm to do so west of Norway. Only when it enters the cyclonic gyre of the Greenland Sea has it been sufficiently cooled to allow its transformation into deep water.

The traditional deep-water circulation scheme [Helland-Hansen and Nansen, 1909] holds that deep water originates in winter in the Greenland Sea, where the densest isopycnals are found close to the sea surface. This source may to some extent be supplemented by cooling and freezing in the Barents Sea [Nansen, 1906]. The deep waters then spread throughout the AM, constrained in doing so only by bathymetry. As the deep waters circulate, they may be modified, especially in passages or over ridges. The principal observational basis for this scheme is that the winter properties of the surface and deep waters are nearly the same in the Greenland Sea, and that the deep waters are colder there than anywhere else in the AM. Present understanding is that the Greenland Sea is indeed a major deep-water formation area, with a renewal time of order 30-40 years [Carmack and Aagaard, 1973; Peterson and Rooth, 1976]. The details of the renewal process are only speculative, with leading candidates being intense convection in small regions, the so-called chimneys [Killworth, 1979], and subsurface formation involving double-diffusive mixing [Carmack and Aagaard, 1973; McDougall, 1983]. Regardless of the formation mechanism, the end product, viz., Greenland Sea Deep Water (GSDW), is the coldest and least saline (\( T < -1^\circ \), \( 34.88 < S < 34.90 \)) of the AM deep-water masses. In the traditional deep-water circulation scheme, the other basins of the AM do not contribute to the deep ventilation.

The first serious modification of this scheme came with the discovery that the dense overflow through Denmark Strait into the North Atlantic consists primarily of intermediate water formed in the Iceland Sea [Peterson and Rooth, 1976; Swift et al., 1980], However, it has recently become apparent from the salinity distribution that deep-water renewal must also occur within the Arctic Ocean [Aagaard, 1981], and it is upon this theme that we focus our attention.

**Deep-Water Formation in the Arctic Ocean**

To provide an overview of the water characteristics of the AM, we have extended into the Arctic Ocean the unpublished meridional section of Swift, Reid, and Clarke through the Greenland and Norwegian seas (Figure 3, section location in Figure 1). Some of the Arctic Ocean stations are actually carefully chosen statistical reductions from a larger, noisier data set assembled from the ice camp stations, such as Alpha in 1957-1958, and from shipborne work near Fram Strait, such as the 1980 Ymer cruise. Other Arctic Ocean stations, such as the 1979 Lomonosov Ridge Experiment (LOREX) profiles, were used intact. The Norwegian and Greenland Sea stations were occupied by the Hudson in March 1982. Because of sparse data and uncertainties in accuracy, the Arctic Ocean portion of this section is not resolved and defined at a level commensurate with that of the southern portion (cf. the appendix for discussion of data quality and uncertainties).

Figure 3a shows that surface waters are warmest in the Norwegian Sea, becoming much cooler in the central Greenland Sea. The small area warmer than 1°C at the surface in Fram Strait represents water from the Norwegian Sea passing westward across the section as it recirculates and moves southward with the East Greenland Current. Relatively warm water also enters the Eurasian Basin, where it sinks beneath the much less dense, cold, but low-salinity polar waters and is gradually modified. In this manner, an intermediate temperature maximum is maintained throughout the Arctic Ocean. This water is normally referred to as the Atlantic layer of the Arctic Ocean, but within the overall context of the water masses of the AM it most nearly resembles the intermediate waters of the cyclonic gyres farther south (the arctic domain in the Swift and Aagaard [1981] terminology), and in the Greenland and Iceland seas it would in fact be defined as a component of Arctic Intermediate Water. In general, when intermediate-depth temperature or salinity extrema are found in water less dense than \( \sigma_t = 32.785 \) (our division between intermediate and deep waters), they are properly referred to as Arctic Intermediate Water, since thermohaline processes in the arctic domain play a major role in their formation. For historical reasons, we are comfortable in continuing to refer to the warmest water in the Arctic Ocean as the Atlantic layer, but we recommend that analogous terminology (e.g., Atlantic Water) not be used in water mass classifications.

Figure 3a also shows that while deep water is much the coldest in the Greenland Sea, it is warmest in the Canadian Basin because the Lomonosov Ridge excludes the coldest deep water. Note, however, that even the Eurasian Basin deep water is warmer than that of the Norwegian Sea, which has been supposed to be its source.

The salinity distribution in Figure 3b suggests many of the same circulation features as does that of temperature. For example, the salinity maximum in Fram Strait is a product of the recirculation of warm and saline water westward. However, our chief interest is in the deep salinities, which are lowest in the Greenland Sea. Since this is also the coldest deep water, so that the GSDW represents an extremum in water mass characteristics, the Greenland Sea must be a source of deep water. However, this source accounts for the deep characteristics only within the Greenland Sea, which contains less than 12% of the deep water of the AM. Deep salinities are of intermediate value in the Norwegian Sea, increase in the Eurasian Basin, and reach a maximum in the Canadian Basin. While the Arctic Ocean salinity values are probably not atypical of the same quality as those to the south, we believe that measurement uncertainties are not such as to significantly distort the distribution shown in Figure 3b (cf. the appendix for a detailed discussion of data quality).

We therefore distinguish between four deep-water masses in
the AM: cold, relatively fresh Greenland Sea Deep Water (GSDW), somewhat warmer and saltier Norwegian Sea Deep Water (NSDW), still warmer and saltier Eurasian Basin Deep Water (EBDW), and the most saline, but much warmer Canadian Basin Deep Water (CBDW).

The upper-layer densities are most easily compared using $\sigma_s$ (Figure 3c). However, because of the temperature dependence of the compressibility, the deep waters should be compared at some greater pressure. In Figure 3d we therefore show the density relative to 3000 dbar. When the relative compressibilities are thus taken into account, EBDW and NSDW are nearly equally dense but are slightly less dense than GSDW. At this pressure, CBDW is by far the least dense deep water. Conversely, at sea surface pressure, GSDW would be the least dense form of deep water.

We now consider the LOREX data [Moore et al., 1983] from the Canadian Basin, together with the deepest values from the Eurasian Basin, as shown in Figure 4. It is clear that not only does the Canadian Basin contain the most saline deep water in the AM (Figure 3b), but also that within the Canadian Basin the salinity is significantly greater below 2000 m than anywhere else in the water column, including the core of the Atlantic layer. Examination of other data sets, generally shallower, confirms the existence of a monotonic salinity increase with depth in the Canadian Basin. The deep salinity is therefore a maximum in both the horizontal and vertical planes, and as pointed out by Aagaard [1981], the conclusion must be that the salt source for the deep water lies neither higher in the water column nor in the nearby deep basins. This contrasts with the hypothesized origin of the Arctic Ocean deep water in the Norwegian Sea [Metcalf, 1960] or as involving a mixing near Fram Strait with water from the Atlantic layer [Timofeyev, 1960]. There is in fact only one likely salt source, viz., the adjacent continental shelf seas, where brine expulsion during freezing produces cold and saline water. This process and its likely importance to the regional salt budget have recently been documented for the northern Bering Sea [Schumacher et al., 1983]. To date the primary interest in the effects of brine formation on the shelves has attached to its role in maintaining the Arctic Ocean halocline [Aagaard et al., 1981; Moore et al., 1983], but we shall show here that the process is probably also in part responsible for ventilating the deep waters of the Arctic Ocean.

In the following discussion we implicitly assume the deep salinity distributions to be steady over the basin renewal times. If the renewal times of the Canadian and Eurasian basins are substantially different (as is probable), we cannot discount the possibility that the higher salinity of the Cana-
Fig. 4. Correlations of θ/S and Si/S near the Lomonosov Ridge. Data from LOREX [Moore et al., 1983]. Inset shows hypothetical mixing lines for intermediate and cold shelf waters to produce deep water; circled points in inset are estimated shelf values. Characteristics of Greenland and Norwegian Sea deep waters are shown by the large solid circle (GSDW θ/S), large solid triangle (GSDW Si/S), and cross (NSDW θ/S).

The Canadian Basin in part reflects the different deep conditions of an earlier period. We are, however, able to make an argument consistent with all available data, based on the simpler steady state assumption. Furthermore, there is at present no realistic basis for dealing with the nonequilibrium case (for further remarks on this, cf. the discussion section).

Based on our inability to find other sources of salt for maintaining the salinity maximum of the deep Canadian Basin, we begin with the assumption that shelf water at the freezing point constitutes the primary salt source for the deep water. (We shall later show evidence for the existence of such water on the shelf.) Examination of the θ/S correlation in Figure 4 then suggests that CBDW is a nearly linear mixture of the freezing point water with water at intermediate depths (near 800 m) in the volume ratio 1:2. The cold shelf water is inferred to have a salinity near 35.1. We shall use this as our working hypothesis and examine it both for consistency and implications.

Figure 4 also shows the silicate/salinity correlation for the same data set. If we extrapolate the correlation to the freezing point salinity suggested by the θ/S correlation, viz., 35.1, we project the shelf source to have a silicate value of about 25 μM L⁻¹. Furthermore, and perhaps more important, the silicate values on the mixing line are appropriate to a 1:2 mixture of freezing point water with water from about 800 m; i.e., the mixing hypothesis based on θ/S properties is also consistent with the silicate data. Aagaard et al. [1981] have argued that the Chukchi and northern Bering seas are likely sources of cold and saline water for the Canadian basin, and although their attention was directed to the halocline, their case can also be extended to deep-water sources. The projected silicate concentration of the shelf source for deep water (~25 μM...
Intermediate Water (AIW) in water mass terminology). If this closely related to the Atlantic layer core (a form of Arctic some extent also its O/S properties, suggest that this water is nature maximum, but its low silicate concentration, and to this water as a parent type in our mixing scheme. We do not in the O/S correlation suggest that it is appropriate to treat feeding the abyss is formed only on the shelves adjoining the Canadian Basin, rather than in a more circumpolar fashion. Indeed, Aagaard et al. [1981] and Swift et al. [1983] have all argued that the Barents and Kara seas have a particularly great potential for forming very saline shelf water. Why, then, is the EBDW not more saline? The answer appears to lie in its connection to the Greenland Sea, whence cold, low-salinity, low-silicate deep water has been observed to enter the Arctic Ocean through the deep channel in eastern Fram Strait [Swift et al., 1983]. Because of the temperature effect on compressibility, GSDW (θ = -1.28øC, S = 34.89) is denser below 1800 dbar than EBDW (θ = -0.85øC, S = 34.94 at the LOREX site, Figure 4). This pressure appears to lie below the Lomonosov Ridge sill depth, so that GSDW is excluded from the Canadian Basin. The GSDW is therefore able to keep the EBDW colder, less saline, and lower in silicate than it would otherwise be, but has little or no effect on the deep Canadian Basin. With respect to silicate concentrations, note particularly that they are identical in the deep water in the Canadian and Eurasian basins down to about 1800 m (Figure 5), but that below this depth the Eurasian Basin values fall off. This is consonant with dilution by GSDW in the deeper waters, but unfortunately our knowledge of the processes affecting the silicate distribution in the Arctic Ocean is insufficient to take the matter further.

We turn now to the evidence for the formation of very dense shelf waters adjacent to the Canadian Basin. Schumacher et al. [1983] reported a maximum daily mean salinity of 35.1 at a mooring in the polynya south of Saint Lawrence in the northern Bering Sea during the 1980-1981 winter, but monthly mean values were much lower (near 33). The formation of dense water was in fact strongly episodic, being forced by the local winds driving a coastal ice divergence.

During winter 1982 we ran a series of conductivity, temperature, and depth (CTD) sections normal to the Chukchi Sea coast of Alaska. The section seaward from Point Lay (Figure 6, location shown in Figure 1) showed salinities within about 20 km of the coast to exceed 36.5, and the 35 isohaline extended seaward at least 40 km in a near-bottom layer. However, other sections showed considerably lower salinities, sug-

![Figure 5](image-url) Vertical profiles of silicate near the Lomonosov Ridge. Data from LOREX [Moore et al., 1983].

![Figure 6](image-url) Salinity section across the northeastern Chukchi Sea, March 3-4, 1982. A data collection gap of 1 day is indicated near the 100-km mark. Section location shown in Figure 1. The water was everywhere at the freezing temperature.
gusting strong temporal variability in the production and northward flow of saline water. The salinity and temperature recorded at mooring CS-2A in Barrow Canyon (location shown in Figure 1) show the variability very clearly (Figure 7), with the temperature following the freezing point as the salinity of the water changed. Two other moorings were also in place in Barrow Canyon during the 1981–1982 winter, and together the set of three moorings provides a velocity section across the canyon. Record-long means from these instruments ranged from more than 20 cm s\(^{-1}\) on the eastern side of the canyon to less than 10 cm s\(^{-1}\) on the western side, but during the last week in February the mean velocity in mid-canyon was about 45 cm s\(^{-1}\). Two winter CDT sections taken across the canyon suggest that the highest salinities were restricted to a layer about 15 m thick and 25 km wide. If we then consider the last week in February, during which the mean salinity at CS-2A was 35.5 (Figure 7), the annual rate of outflow this saline through Barrow Canyon was 3.2 × 10\(^3\) m\(^3\) s\(^{-1}\) (0.003 sverdups (Sv)). If this water mixed with water below the temperature maximum (\(S = 34.89\)), this shelf source alone could renew the CBDW in about 2000 years.

Apart from the fallacy of annualizing an event in this manner, such calculations are obviously also sensitive to the mixing ratio used, i.e., to the salinity both of the cold shelf water and of the ambient water. For example, if this same shelf water mixed with ambient water of 34.92 (found at about 1300 m in the LOREX data), a 500-year residence time would require an off-shelf flux of 10\(^4\) m\(^3\) s\(^{-1}\) (0.01 Sv), whereas shelf water of 35.1 mixing with ambient water of 34.89 would require a 0.06 Sv flow. The point is that extremely saline water is clearly found on the shelf during winter, but that uncertainties both in the formation rates (which might vary greatly from year to year) and in the subsequent mixing processes do not presently allow a meaningful residence time to be calculated for the CBDW. Such an estimate is probably most easily had from appropriate tracer measurements, e.g., \(^{14}\)C. When such measurements become available, they will likely indicate a large difference in the residence times of the deep waters on either side of the Lomonosov Ridge, since the deep Canadian Basin appears primarily to be ventilated by relatively small volumes of shelf water (hence the relative warmth of the CBDW), whereas the Eurasian Basin is also ventilated by the influx of GSDW, for which the renewal time in the Greenland Sea is only 30–40 years [Carmack and Aagaard, 1973; Peterson and Rooth, 1976].

In the Eurasian Basin an additional shelf source of dense water is indicated, viz., the cooling of water already saline by virtue of its being derived from the northward flow of Atlantic water in the Norwegian Sea. Such water does not require its salt augmented through freezing in order to become denser than EBDW and is in this respect distinct in its origins from the shelf waters discussed earlier, although obviously hybrids of these processes are conceivable. The Barents Sea is a particularly likely site for the cooling of saline water, since it receives a considerable inflow of Atlantic water from the Norwegian Sea. Such a scheme was proposed by Swift et al. [1983], but while they cited observations of dense water in the Barents Sea (Figure 8), the salinity of the water colder than 0° C was less than 34.94. Certainly the potential for forming exceedingly dense water in this area exists, and L. Midttun (unpublished document, 1984) has recently reported water near −1°C and more saline than 34.95 on the southern side of the Novaya Zemlya–Franz Josef Land passage. From there such water might enter the Arctic Ocean either via the northeastern Barents Sea or through the Kara Sea.

Our conclusion is therefore that deep water is formed within the Arctic Ocean itself, including the Canadian Basin, but that the critical observations regarding rates, sites, and the relative importance of the several probable mechanisms remain to be accomplished.

**OUTFLOW OF DEEP WATER FROM THE ARCTIC OCEAN**

We have argued that shelf waters which primarily acquire their great density from the addition of brine during freezing, mix with water from below the temperature maximum in the Arctic Ocean to form new deep water. This deep water is more saline, warmer, and higher in silicate than the GSDW, and we shall look for such a signature in Fram Strait, where we might expect a juxtaposition of the water masses if there is significant outflow of deep water from the Arctic Ocean.

In this context, the higher-order effects of pressure on density become important. In general, there exists some pressure
at which the in situ densities of two water types of slightly different temperature and salinity are equal. We define this as the compensation pressure, \( p_c \). For \( p < p_c \), the in situ density of the colder, less saline water is less than that of the warmer, more saline water, but for \( p > p_c \), the colder water is denser. The importance of this differential compressibility effect in the stratification of the deep ocean was apparently first pointed out by Ekman [1934]. Consider now the situation in which two very deep isothermal and isohaline water columns are separated by a vertical partition, one column being colder and fresher than the other, and assume that \( p_c \) exists within the range of pressure exhibited by these columns (Figures 9a and 9b). If the partition is removed, the warmer and saltier column will collapse about the level of \( p_c \) to form a temperature and salinity maximum at middepth (Figure 9c). If we take the warmer water to represent EBDW entering Fram Strait and the colder water to represent the resident GSDW, and at the same time consider the effect of rotation, we should expect to find the outflowing core of EBDW trapped against the western boundary near the level \( p_c \). The core is constrained in the vertical plane by the compressibility terms in the equation of state and in the horizontal plane by rotational effects. For \( \theta/S \) values of the EBDW near Fram Strait of \(-0.95^\circ C, 34.93\), and of the GSDW of \(-1.28^\circ C, 34.89\), \( p_c \) is in the range 1900–2000 dbar, equivalent to about 1900 m.

Recent data show that there is in fact an outflow of EBDW to the Greenland Sea and that it behaves in the manner suggested by the above arguments. Figure 10 shows a composite salinity section drawn from the 1982 Meteor observations in the western Greenland Sea. The data were kindly made available by K. P. Koltermann. The core of saline water (also relatively warm and silica-rich) against the Greenland slope
near 1500 m can only represent deep outflow from the Arctic Ocean. A similar situation is also suggested by the Hudson data, and one can follow this outflow as it is incorporated into the periphery of the convectively renewed cyclonic gyre of the Greenland Sea and carried southward. In the process, the salty outflow is freshened by the convective products of the Greenland gyre, and the modified deep water flows into the Norwegian Sea through gaps in the mid-ocean ridge northeast of Jan Mayen to ventilate the southern basins of the AM. In this connection, it is probable that the compensation pressure again plays a special role. Other factors being equal, we should expect the most effective mixing between the two water types to occur at the level of \( p_c \), since the stratification disappears there.

A further aspect of this situation is that the compressibility effect alone could conceivably sustain a geostrophic shear. Consider again the two water columns of Figure 9. The difference between the two columns of the vertical integral of pressure, \( \Delta p_{\rho_c} \) (Figure 9d), increases with depth to a maximum at \( p_c \) and then decreases with increasing depth. The relative geostrophic motion associated with this horizontal pressure gradient has a maximum at \( p_c \) and is directed out of the page.

The internal deformation radius, with stratification based solely on the differential compressibility effect, is very small, about 1.4 km for the \( \theta/S \) values cited near Fram Strait. This length scale gives a current core near 10 cm s\(^{-1}\). However, the data of Figure 10 suggest an actual core width of at least 10–20 km, requiring a proportionate reduction in baroclinic velocity.

**ROLE OF THE INTERMEDIATE WATERS**

We have focused upon the formation and circulation of the deep waters of the AM because these represent the most extreme reaches of the ventilating processes. However, these deep waters primarily circulate internally within the AM, since they lie below the greatest densities in free communication with the North Atlantic. The only pathway of escape for the densest waters is through vertical mixing with the overlying layers. In contrast, waters of lesser density are produced near the sea surface in all the basins and found at intermediate depth, and certain of these intermediate waters pass over the Greenland-Scotland ridges into the North Atlantic.

The intermediate waters have their principal origin at the surface of the Iceland and Greenland Sea gyres, where winter convection creates thick layers of cold and well-oxygenated water. This water has been called upper Arctic Intermediate Water (upper AIW) by Swift and Aagaard [1981]. As this water intrudes into the Norwegian Sea, it forms an intermediate salinity minimum (Figure 3b) and also an oxygen maximum. The AIW is also carried northward through eastern Fram Strait, where it mixes with the warm waters of the West Spitsbergen Current, thereby decreasing the temperature of the warm inflow to the Arctic Ocean.

Within the Arctic Ocean, outflows from the peripheral shelf seas probably mix with the Atlantic layer (AIW in water mass terms), continually increasing the density of the Atlantic layer as it circulates. This is entirely consistent with the modification of the Atlantic core observed by Coachman and Barnes [1963], and in fact, Aagaard et al. [1981] calculated the shelf water outflow in such a scheme to have a mean salinity near 34.7.

The intermediate waters are a critical component of the thermohaline circulation in the AM, since their density surfaces generally outcrop in winter, and the waters are thus relatively well ventilated. Their annual formation provides a rapid replenishment of a set of rather small reservoirs [Swift and Aagaard, 1981], the contents of which are subsequently modified during spreading and mixing. Not only does this give rise to a rather broad range of water properties, but the short replacement times for the intermediate waters make them very responsive to environmental changes at the sea surface. These changes can be transmitted quickly (within 2–10 years) both within the AM and also to the North Atlantic [Gordienko et al., 1976; Brewer et al., 1983; Swift, 1984]. The connection between the ventilating regions and the deep North Atlantic is most nearly direct through Denmark Strait, and this path and the associated time scales have been discussed by Swift et al. [1980] and by Livingston et al. [1985]. The other primary outflow to the North Atlantic, through the Faeroe Bank Channel, is derived from a somewhat deeper level than that through Denmark Strait. The Hudson data suggest that a blend of intermediate water recently ventilated in the gyres with deep water (which must mix upward into the intermediate waters as it is displaced by new deep water) seems to account best for the properties of the outflow through the Faeroe Bank Channel. This outflow is less sensitive than that through Denmark Strait to environmental changes induced at the sea surface [Brewer et al., 1983; Swift, 1984; Livingston et al., 1985], undoubtedly because it is more strongly mixed with the deeper layers in the Norwegian Sea.

**DISCUSSION**

The intention of this paper has been to synthesize the deep and intermediate circulation of the AM in a manner consistent with recent data. We conclude that to explain property distributions adequately, there must be at least two major sources of deep water: open-ocean convection in the Greenland Sea, and near-boundary convection around the perimeter of the Arctic Ocean. The various deep-water varieties, with salinities between 34.89 and 34.96, all lie well below the southern sills which confine the AM system, and the deep waters thus are constrained to circulate internally. In the Iceland Sea in particular, open-ocean convection penetrates only to middepth, and the resultant intermediate water, together with similar water from the Greenland Sea, comprises the bulk of the overflow into the North Atlantic.

We have summarized our synthesis schematically in Figure 11. In a plane view (Figure 11a) the major circulation features of at least the upper ocean, many of which are primarily wind driven, are the Bering Strait inflow (A), the Beaufort gyre (B), the transpolar drift (C), the East Greenland Current (D), the Norwegian Atlantic Current (E), the West Spitsbergen Current (F), the Greenland gyre (G), and the Iceland gyre (H). In a vertical section along the 180°–0° meridians (Figure 11b), open-ocean convection is shown to reach the bottom in the Greenland gyre, while in the Iceland Sea such convection reaches only to middepth. In the Arctic Ocean, near-boundary convection renews water masses down to the bottom of the basin. The selected isopycnals in Figure 11b separate surface, intermediate, and deep waters. The water mass classification appropriate to this circulation scheme (Figure 11c) largely follows Swift and Aagaard [1981] but further expands the deep-water classification.

We do not propose that the distributions of water properties represent a strict equilibrium. Indeed, it is clear that the exchanges with the North Atlantic exhibit \( \theta/S \) variations on the decadal time scale [Swift, 1984]. There is also evidence of
significant temperature changes in the GSDW over only a few years [Aagaard, 1968]. We consider it likely, however, that while the balance of the thermohaline processes may shift from time to time, and thereby produce T/S variations in the various water masses, the basic features of the thermohaline circulation portrayed in Figure 11 have not changed significantly in recent times. We further believe that the patterns of interbasin exchange shown in Figure 11 are well supported by the data presented in this paper and elsewhere in the literature [e.g., Aagaard, 1981; Swift et al., 1983]. In particular, we note the role of the Lomonosov Ridge in restricting the GSDW from freshening the deep Canadian Basin. This suggests that the Canadian Basin was also in the past more saline than the Eurasian Basin. However, most of the details of the production of deep and intermediate waters required to drive the circulation suggested by Figure 11 are not well established.
Fig. 12. Schematic representation of a stratified sinking plume at the freezing temperature. The sequential shaving off and interleaving of successively denser layers modifies both the plume and the ambient \( \Theta/S \) curves as in the insets on the right-hand side.

For example, while there is general agreement that deep convection and ventilation occurs at least intermittently in the Greenland Sea [Carmack and Aagaard, 1973; Swift and Aagaard, 1981], the overturn process itself has never been observed [Killworth, 1979].

An essential supposition in the conceptual model we have suggested is that dense water forms on the continental shelves surrounding the Arctic Ocean and that this water sinks to depth in near-boundary convection. It is further likely that deep-water ventilation occurs in this manner in both the Canadian and Eurasian basins, although in the Eurasian Basin renewal from the Greenland Sea is probably also an important process. While there is ample evidence that shelf water of sufficiently high salinity (density) to displace deep water is formed in winter, obviously missing are observations of convection along the continental slope itself. The situation is analogous to that pertaining to bottom water renewal around Antarctica prior to the International Weddell Sea Expeditions in the late 1960's and early 1970's [cf. Foster and Carmack, 1976]. A similar observational program is probably required for the Arctic.

**Implications for Modeling**

A recent review of modeling of the Arctic Ocean [Killworth and Carmack, 1983] included two areas particularly relevant to the present discussion: vertical TS structure and turbulent plume dynamics.

Three recent papers have attempted to explain the basic TS structure. Stigebrandt [1981] used a two-layer model to examine the combined effects of river runoff, basin inflow, and ice formation on surface water characteristics; because of the two-layer approximation, no account was taken of deep water. A more detailed examination of water mass structure and circulation was made in Semtner's [1976] three-dimensional simu-
lution. However, this model used as boundary conditions inflow/outflow values published prior to 1973. In particular, the model included neither an outflow of deep water from the Arctic Ocean nor the production of dense shelf waters, which we feel to be of critical importance.

Killworth and Smith [1984] constructed a filling-box model of the Arctic Ocean to examine the role of shelf-derived plumes in maintaining the vertical TS structure of the cold halocline. In this model, horizontal mixing from the sides was supposed to be instantaneous, and inflowing water masses were required to spread adiabatically at the depth at which their in situ density matched that of ambient basin water. Shelf water itself was assumed to be horizontally uniform and to sink without mixing to its equilibrium depth. Although this model yielded an adequate simulation of the surface waters, it failed to reproduce the TS structure below about 200 m. We believe the discrepancy may be reconciled by the ideas we have presented. In a filling-box model the key dynamical variable is the vertical velocity forced by continuity at the level at which a plume enters the interior and raises overlying water. Indeed, the poor simulation below the halocline led Killworth and Smith to wonder if their formulation had omitted an important process, e.g., deeper plume penetration which would alter the temperature, salinity, and vertical velocity fields in the deep water.

We suggest that the filling-box approach provides an especially useful framework for viewing the water mass structure of the AM, and that means to be found to include the ideas presented here. Specifically, account should be taken of shelf water, or more probably its derivatives, penetrating to the bottom, and provision should be made both for nonuniform shelf water sources and for appropriate mixing with ambient waters during sinking.

The sinking of turbulent plumes along a sloping bottom in polar seas has been treated by Killworth [1977], Carmack and Killworth [1978], and Melling and Lewis [1982], using the streamtube formulation of Smith [1975]. While such models yield good insight into the relative importance of buoyancy, entrainment, and bottom friction, they are limited in their ability to explain water mass formation by the assumption that at any position along its path the plume is homogeneous. Therefore in any given numerical experiment, spreading into the interior can only occur at a single depth.

By contrast, there is evidence that at least some sinking plumes are stratified [Foster and Middleton, 1980; Thorpe, 1984; Carmack, 1985]. Now, in a stratified plume a spectrum of TS points is being transported downward. Since a given water type cannot be carried below its in situ density level (except by penetrative convection), it follows that for such a plume, detachment into the interior will occur at a variety of depths along the continental margin. We will tentatively refer to this process, by which the buoyancy of ambient waters lifts off successive layers of the plume, as "shaving" (Figure 12), and we suggest that a means of parameterizing this process be developed. We deem it likely that such stratified plumes have both an upper stratified portion and a lower well-mixed portion. At any level it would be the upper portion that actually mixes with ambient water. This upper portion of the plume must maintain a density greater than or equal to that of the ambient water and would therefore continually be shaved off to interleave along isopycnal surfaces. Shaving is therefore a form of detrainment, in which elements of a plume are mixed into an ambient fluid of reduced turbulence. The effect of such a process on the TS structure is shown in Figure 12.

Finally, we state our belief that modeling related to the thermohaline circulation in the Arctic is particularly timely, since major observational efforts are still in their infancy. There is therefore ample opportunity for theoretical constructions to guide experimental design.

APPENDIX

Notation

Our hydrographic notation is as follows. Salinity units, which are nominally 10^-3 mass mass^-1, are expressed without the factor of 1000, as we generally use the Practical Salinity Scale (PSS 78). However, older salinities were not re-determined with the new scale. In the salinity range of the deepwaters the difference between the new and old scales is <0.001 and is not significant, nor do our conclusions rest on the small differences between the new and old scales over the much greater surface salinity ranges. We have expressed densities in sigma notation, e.g., \( \sigma_2 = \rho_2 - 1000 \text{ kg m}^{-3} \), where \( \rho_2 \) is the in situ density in kg m\(^{-3}\) the water would have if moved adiabatically to the depth where the pressure is 2000 dbar (= 20 MPa). In computing density we have used the International Equation of State (EOS 80). Had the Knudsen equation of state been used, the "densities" would actually have been specific gravities, and would have been higher, by about 0.02 at the sea surface and 0.05 at 2000 dbar.

Comparison of Density Surfaces

The water found on a given isopycnal has some range in temperature and salinity, and if moved adiabatically to some different reference pressure it would exhibit a range of densities there, rather than a single value. In Figure 2 we have chosen an appropriate reference pressure for each isopycnal used in our analysis. To assist the reader in comparing Figures 2 and 3, we have calculated the approximate range of density surfaces (in \( \sigma \)-notation) at 0, 1000, 2000, and 3000 dbar, appropriate to the actual \( \theta/S \) ranges on each of the isopycnals used in Figure 2 (Table A1).

Data Sources and Errors

Our interpretations of the deep circulation are dependent on small differences in salinity and temperature, and so are sensitive to measurement errors. The data from the Greenland and Norwegian seas used in Figure 3 were collected during Hudson cruise 82-001, February 28 to April 6, 1982, by J. H. Swift, J. L. Reid, and R. A. Clarke. A data report is available from Swift. The deep temperatures are thought to be accurate within 0.01°C, and the deep salinities to about 0.002 with respect to Wormley P90 standard seawater. The deep water masses sampled during the Hudson cruise included not only NSDW and GSDW, but also EBDW in Fram Strait. This

| TABLE A1. In Situ Density Ranges (in \( \sigma \)-Notation) at 0, 1000, 2000, and 3000 dbar of the Water Masses Found on the Three Characteristic Density Surfaces Used in Figure 2 |
|-----------------|-----------------|-----------------|
| Density Surface | \( \sigma_0 = 27.9 \) | \( \sigma_1 = 32.785 \) | \( \sigma_2 = 37.457 \) |
| Pressure, dbar  | 0  | 1000 | 2000 | 3000 |
|                 | 27.9 | 32.52-32.67 | 37.04-37.33 | 41.46-41.88 |
|                 | 28.02-28.07 | 32.785 | 37.39-37.44 | 41.89-41.99 |
latter sampling added much to our confidence in the EBDW characteristics.

The Hudson section was extended north through Fram Strait with CTD-derived temperature and salinity data from the Ymer 1980 cruise [Anderson and Dyrssen, 1980]. Where Ymer coverage overlapped with that of the Hudson (and of the Transient Tracers in the Ocean North Atlantic Study in 1981), the deep salinities were compared. As a result, the Ymer salinities reported by Anderson and Dyrssen [1980] were uniformly reduced by 0.013 to conform with the salinometer-derived values from the other data sets [Swift et al., 1983]. These adjustments in effect represent a recalibration of the Ymer data, assuming that the deep waters are relatively invariant spatially and temporally. Thus adjusted, the Ymer salinities show not only the same relative distribution as those from the Hudson (e.g., an EBDW salinity increase to the bottom and a nearly homohaline NSDW) but also nearly the same numerical values. We note also that unpublished data collected in 1984 north of Fram Strait by Swift on the Polarstern, using the same standards and procedures as those of the Hudson cruise, yield EBDW values consistent with those cited here.

The salinities from the Lomonosov Ridge Experiment (LOREX) were important in shaping our conclusions. These data span both sides of the Lomonosov Ridge and demonstrate striking differences between CBDW and EBDW (Moore et al. [1983]; discussed in greater detail by Aagaard [1981]). The LOREX profiles were collected over 38 days, and the analysis environment was not ideal. This introduced noise into the measurements, but the signals basic to our interpretation were well above the random errors. All indications are that the LOREX deep salinities are sufficiently accurate for our purposes: EBDW values are similar to those from the Fram Strait expeditions, and Canadian Basin values are very close to those measured by R. M. Moore (unpublished data, 1983) during the 1983 Cesar expedition near 86°N, 110°W. We were thus encouraged in our efforts to produce a basin-scale salinity intercomparison extending over the whole of the AM.

We are not so confident in the quality of the remaining data sets used in Figure 3, distilled from a search of National Oceanographic Data Center (NODC) files. Individual profiles were often noisy and occasionally contained glaring inaccuracies. If these were our only guides, we would have been restricted to a speculative paper. We do note that after eliminating obvious errors and averaging horizontally and vertically, these data usually showed features consistent with the results from our principal data sets, and so were a useful aide in contouring the section across the deep basins. Because our conclusions do not rest upon these subsidiary data, we will not discuss them further.

We also note that the coverage of Meteor cruise 61 (data used in Figure 10) overlapped with that of Hudson cruise 82-001 (which took place 2 months earlier). An intercomparison of deep-water temperatures and salinities showed on the average the meteor samples were warmer by 0.003°C and fresher by 0.002 than equivalent samples from the Hudson. These systematic differences are much smaller than the EBDW outflow signal shown in Figure 10.

Finally, we consider the winter Chukchi Sea data and, in particular, the salinities. The observations for Figure 6 were made with a Neil Brown Instrument Systems Mark IIIIB CTD system carried by helicopter; the underwater unit had been reconfigured to allow deployment through an 8-in. auger hole in the ice. All sensors were laboratory calibrated prior to field deployment. Bottle calibrations in the field were extremely difficult during the Chukchi work in 1982 because the helicopter was nearly unheated, so that the sampling bottle could rarely be kept from freezing and leaking. (The water sample was already at the freezing point, and the actual cast was made a short distance away from the helicopter in an unprotected location.) On the Point Lay section (Figure 6), only one successful bottle sample was obtained, and unfortunately, it was located in the steep salinity gradient near bottom and is therefore of limited usefulness in calibration. However, on the section immediately south of the Point Lay section, a good calibration was obtained 2 days earlier, when a sample bottle was kept from freezing and leaking and three subsamples agreed with the CTD to within ±0.002 in salinity. A review of the entire data set with respect to the laboratory calibrations, the very few field calibrations, internal consistency in the TS distributions, and maintenance of the freezing temperature as a function of salinity suggest that the salinity determinations are accurate to within at least 0.02. This is ample to ascertain the presence of the anomalously saline water, which approaches 4 above the far-field salinity.

With respect to the serial data of Figure 7, the only meaningful calibration was that performed in the laboratory. A CTD station taken 11 km away, but in nearly the same water depth, showed the identical temperature with that of the moored sensor, but a salinity 0.37 lower. A second CTD station 16 km away and somewhat shallower, recorded the same temperature to within 0.015°C; the CTD salinity was only 0.04 lower than at the mooring, and 1 m from the bottom the CTD showed a salinity 0.02 higher. The minimum temperature recorded at the mooring during the winter was −1.97°C, which is identical with the freezing point at the salinity calculated for the simultaneously recorded conductivity, viz., 35.84. We therefore accept the mooring data based on the laboratory calibration but recognize that an absolute salinity error of as much as a few tenths is possible. We note, however, that the temperature in Figure 7 follows the freezing point very closely, suggesting that errors in the record are not large.

Acknowledgments. Among the many colleagues whose assistance was indispensable were R. A. Clarke and J. L. Reid in collecting the Hudson data; D. J. Hanzlick, S. D. Harding, and R. B. Tripp, the Chukchi Sea data; K. P. Koltermann, the Meteor data; M. G. Lowings and R. M. Moore the LOREX data; and B. Rudels the Ymer data. E. A. Aagaard performed the hypsographic calculations, and M. Mitchell and J. Washington constructed many of the figures. The work was supported by the National Science Foundation through grant DPP-8100153 and the Office of Naval Research through contracts N00014-75-C-0893 and N00014-84-C-0111 to the University of Washington; by the National Science Foundation through grant OCE-8320410 and the Office of Naval Research through contract N00014-80-C-0440 to the Scripps Institution of Oceanography; and by the Marine Minerals Service through interagency agreement with the National Oceanic and Atmospheric Administration, as part of the Outer Continental Shelf Environmental Assessment Program. This is contribution 1403 from the School of Oceanography, University of Washington, and also constitutes a contribution from the Marine Life Research Group, Scripps Institution of Oceanography.

References


K. Aagaard, School of Oceanography, University of Washington, Seattle, WA 98195.

E. C. Carmack, Department of the Environment, NWRI Branch, 4160 Marine Drive, West Vancouver, British Columbia, V7V 1N6 Canada.

J. H. Swift, Scripps Institution of Oceanography, MLR Group, A-030, La Jolla, CA 92039.

(Received August 17, 1984; accepted September 13, 1984.)