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Water mass components of the North Atlantic deep western boundary current

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Abstract—Four hydrographic sections across the North Atlantic deep western boundary current from 55°W to 70°W are analysed to distinguish the current's different water mass components. The deepest component is the Norwegian–Greenland overflow water (2–3°C) which is characterized most readily by a core of high oxygen, tritium, chlorofluorocarbons (CFCs) and low silicate anomaly. The above lying Labrador Sea Water (3–4°C) is distinguishable at this latitude only by its core of low potential vorticity. The shallowest component of the boundary current (4–5°C) is revealed by a core of high tritium, CFCs and low salinity anomaly but has no corresponding oxygen signal because of its proximity to the pronounced oxygen minimum layer. A careful analysis of the shallow water mass reveals that it is not dense enough to be formed in the central Labrador Sea even during warm winters. Rather, based on historical hydrography its area of formation is the southern Labrador Sea inshore of the North Atlantic current where surface layer salinities are particularly low. A simple scale analysis shows that lateral mixing with the adjacent North Atlantic current can increase the salinity of this component to the values observed in the mid-latitude data set.

INTRODUCTION

TRACERS have long been used in oceanography to help describe the state of the ocean, and they continue to be an important contributor in the on-going effort to understand the ocean's general circulation. Early oceanographers deduced considerable information regarding ocean circulation by interpreting lateral and vertical distributions of temperature, salinity and dissolved oxygen. In recent years, improved technology has allowed us to measure a whole new suite of transient tracers now penetrating the world ocean. When different tracers are used in conjunction with each other there are more potential constraints, such as the various tracer based ages. Two common examples of this are the chlorofluorocarbon (CFC) age (e.g. WEISS *et al.*, 1985) and the tritium–helium age (e.g. JENKINS and CLARKE, 1976). In the presence of substantial mixing, however, determining water mass age from transient tracers becomes increasingly difficult (PICKART *et al.*, 1989; THIELE and SARMIENTO, 1990).

Tracers are particularly valuable in the deep ocean which is isolated from atmospheric contact. These abyssal waters are slowly replenished by boundary currents which themselves originate at the sea surface at high latitudes and thus carry anomalous quantities of various properties (i.e. anomalous relative to the neighboring interior water). The present paper focuses on the anomalous nature of the North Atlantic deep western boundary

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current (DWBC), which flows equatorward along the western continental slope from depths as shallow as 700 m to depths as great as 4000 m. Various authors have referred to this entire body of water as the DWBC, while others refer to just the deeper flow (say >2000 m) as the DWBC. In this paper three water masses are distinguished which are referred to as separate components of the DWBC.

The data come from a single hydrographic survey conducted in 1983 consisting of four DWBC crossings from 55°W to 70°W. Altogether there are 21 DWBC stations and 26 stations occupied in the interior basin, with measurements of temperature, salinity and various chemical tracers. Using these data we characterize the different water masses of the DWBC and summarize their distinguishing features, i.e. what are the best tracers of each component and why. The deepest component of the DWBC is the diluted Norwegian–Greenland Sea overflow water (NGOW); above this is the classical Labrador Sea Water (LSW) and shallower yet is a water mass which originates in the southern Labrador Sea. The former two components have been well studied and their formation is fairly well understood. The emphasis of this paper is on the origin and downstream character of the shallowest DWBC component, whose overwhelming distinguishing feature is a core of high CFCs and tritium. This water has been identified in the literature as LSW, although we show that it is not dense enough nor does it have the correct T–S characteristics to be classical LSW.

DATA AND METHODS

Figure 1 shows the station locations of four CTD sections of R.V. *Oceanus* cruise 134 (OC134) occupied in summer of 1983. We have divided each section into two parts, those stations on the continental slope within the DWBC (shallower than about 4200 m) and the remaining interior stations (Fig. 1). To make the analysis as compact as possible, a composite vertical profile of the 21 boundary stations was constructed. First, each station was linearly interpolated onto a regular pressure grid with a grid spacing of 20 db. The average property value at each pressure was then computed (using only those stations with observed data at that pressure), and the resulting profile was Gaussian smoothed with a filter of width a = 150 db (except where noted).

To illuminate the anomalous nature of the DWBC, an anomaly profile was computed relative to the interior stations. First a composite profile of all the interior stations was computed on a potential temperature (θ) grid (i.e. the averaging was done at each temperature level). A temperature grid was used because the interior stations include the Gulf Stream which has strongly sloping isotherms with depth (near the boundary, however, isotherm slopes are weak so a pressure grid suffices there). The anomaly is then defined simply as the difference between the boundary and interior value at a given value of θ. We chose potential temperature as the independent variable instead of potential density so that the profiles could be compared more readily to historical results (it should be noted that θ varies only slightly with potential density across the four OC134 sections). The following tracers were analysed: salinity, oxygen, potential vorticity, silicate, chlorofluorocarbon F-12 and F-11:F-12 ratio, and finally the anomaly of each of these variables.

WATER MASS COMPONENTS

The three components of DWBC correspond roughly to the following temperature intervals (Fig. 2a): from 2–3°C is the NGOW found at depths 2500–4000 m, above this

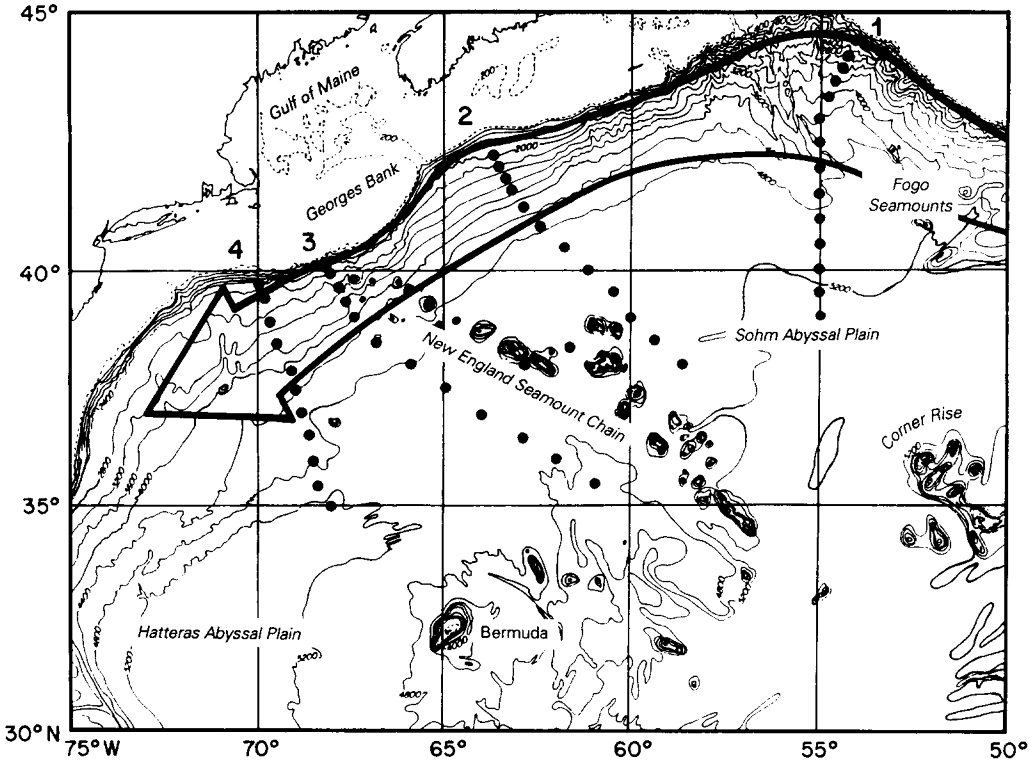

Fig. 1. Station positions of the four hydrographic sections of OC134. The lateral extent of the DWBC tracer signal is indicated by the schematic current enclosing those stations used for the average boundary profile.

(1300–2500 m) is the LSW in the temperature range 3–4°C, and from 4–5°C (700–1300 m) is a water mass which we will simply call shallow DWBC water. A vertical section of F-12 from OC134 (Fig. 2b) shows two distinct cores in the NGOW and shallow DWBC water, respectively. The seaward extent of these cores was used as the division between boundary and interior stations.

The resulting composite vertical profiles are shown in Figs 3–7. Salinity is not included as it warrants special consideration below. The oxygen and potential vorticity profiles required special averaging because of their noisy character. For potential vorticity the raw data were first Gaussian smoothed with a filter of width $a = 90$ db, then the final composite profile was smoothed with $a = 700$ db (compared with 150 db for the other variables). For oxygen the only change was a Gaussian smoothing of the raw data with $a = 50$ db.

Norwegian–Greenland Sea overflow water

The deepest component of the DWBC consists of a combination of waters which are formed in the Norwegian and Greenland Seas. These newly formed deep waters enter the northern North Atlantic through the Faeroe Bank Channel, Iceland–Scotland Ridge and Denmark Strait. Overflow from the former two combine as a single current and flow through the Gibbs fracture zone, eventually encountering the Denmark Strait overflow. It

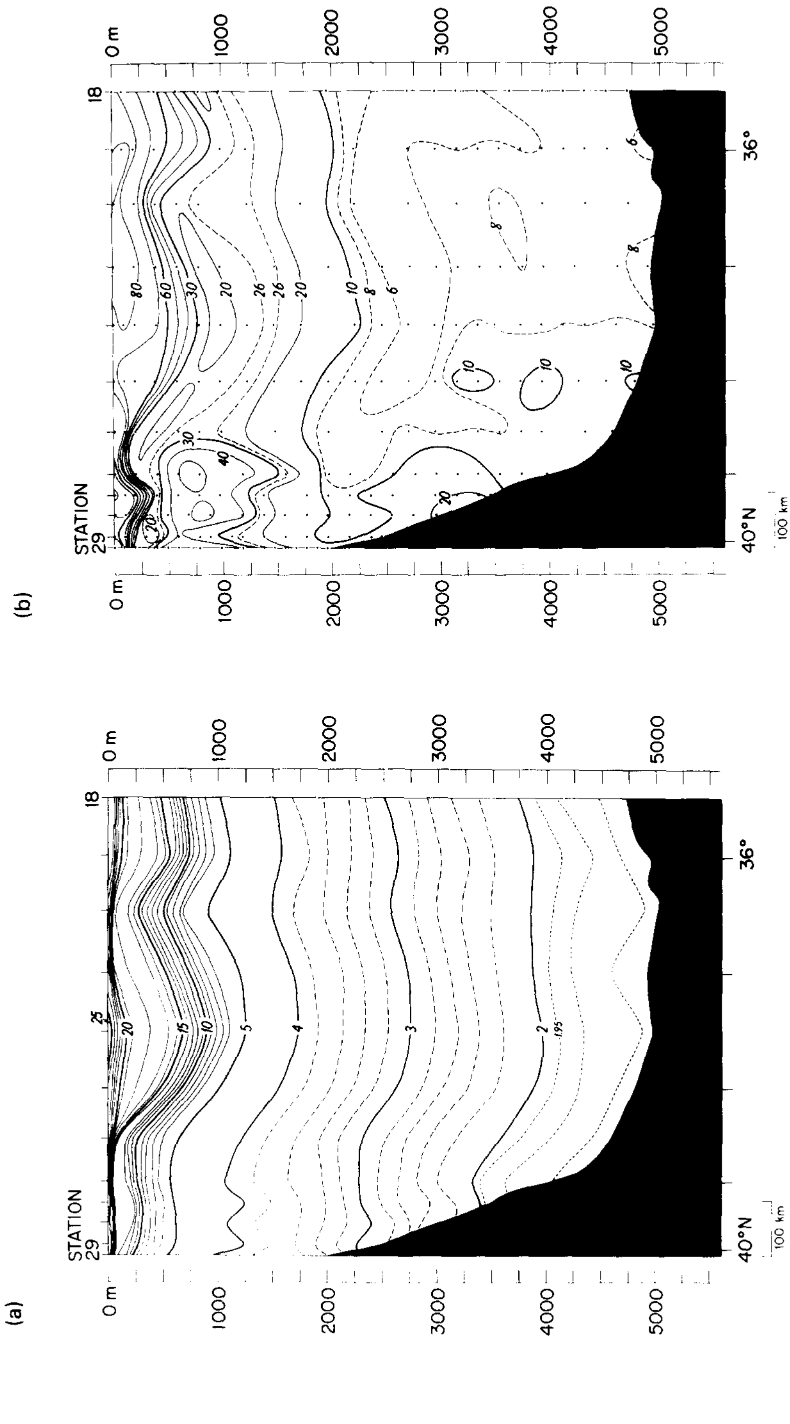

Fig. 2. (a) Potential temperature (°C) of OC134 section 3. (b) Section of F-12 (p mol kg^{-1} × 100).

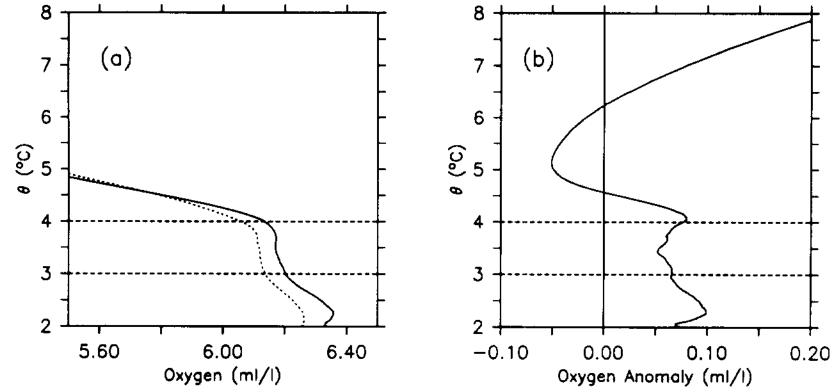

Fig. 3. (a) Average profile (solid line) of oxygen vs potential temperature (colder than 8°C) for all stations in the DWBC (see Fig. 1). The dashed line is the average interior profile. (b) Oxygen anomaly (see text for how anomaly is computed).

is unclear just how the 4–5 Sverdrups (Sv) (Steele *et al.*, 1962) of Gibbs fracture zone water mixes and interacts with the 2–3 Sv (Ross, 1984) of Denmark Strait water, though by the time the combination of waters rounds the Grand Banks it seemingly comprises a single well defined mean flow (Pickart, 1991). Because of contact with the atmosphere prior to formation, this component of the DWBC is easily identified by its elevated oxygen concentration (e.g. Clarke *et al.*, 1980; Fig. 3a). This signal coincides with a core of high CFCs (Fig. 4a) and tritium (Fine and Molinari, 1988; Jenkins and Rhines, 1980; Olson *et al.*, 1986). The NGOW is characterized in the OC134 data set by a core of high CFC ratio as well (Fig. 5a). However, since the atmospheric F-11:F-12 ratio has been constant since 1978, a section occupied today will contain a uniform CFC ratio throughout the DWBC.

The anomalies of oxygen, F-12 and F-11:F-12 ratio (Figs 3b–5b) contain analogous deep cores in the 2–3°C temperature range. The situation is different, however, for silicate. While there is no atmospheric flux of silicate, the near-surface water which sinks to form the deep DWBC is nutrient-poor relative to abyssal interior waters. The composite silicate profile (Fig. 6a), however, shows no deep core, only a change in slope below 3°C. This may

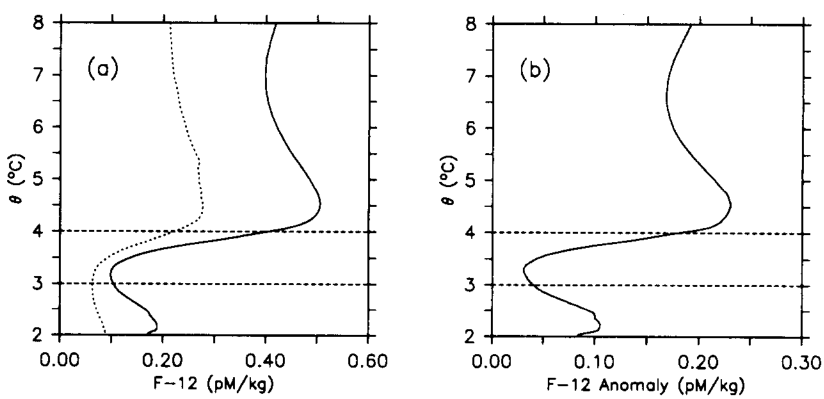

Fig. 4. (a) Average profile of F-12 vs potential temperature (colder than 8°C) for all the stations in the DWBC (see Fig. 1). (b) F-12 anomaly.

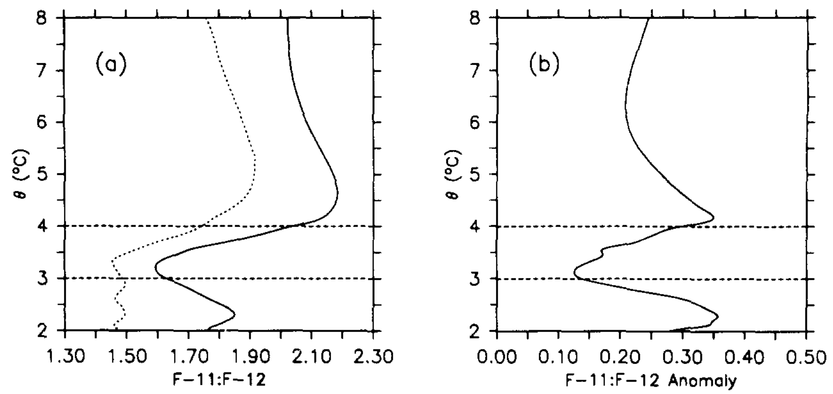

Fig. 5. (a) Average profile of F-11 : F-12 ratio vs potential temperature (colder than 8°C) for all the stations in the DWBC (see Fig. 1). (b) F-11 : F-12 ratio anomaly.

be because the silicate signal is modified more efficiently due to flux from the sediments (e.g. EDMOND *et al.*, 1979) in addition to mixing with adjacent waters, whereas only the latter process reduces the oxygen and CFC cores. It is not until the silicate anomaly profile is computed (Fig. 6b) that the DWBC signature is clearly revealed as an extremum.

Although the merging of Denmark Strait water with Gibbs fracture zone water is not fully understood, most of the NGOW tracer signal comes from the Denmark Strait contribution. SMETHIE and SWIFT (1989) conclude from the tritium–krypton-85 ratio that the time for newly convected water to pass through the Denmark Strait is an order of magnitude shorter than for the Faeroe Bank channel. They argue that the Faeroe Bank channel flow actually receives most of its tracer input from mixing with ambient water after overflowing the sill. LIVINGSTON *et al.* (1985) have shown that the Denmark Strait overflow is also characterized by anomalously high levels of radionuclides.

It is of interest to compare the magnitude of the NGOW tracer cores observed in the OC134 data set with values measured further north (closer to the source). When the core values of oxygen, silicate, salinity and F-11 from a section across the western boundary of

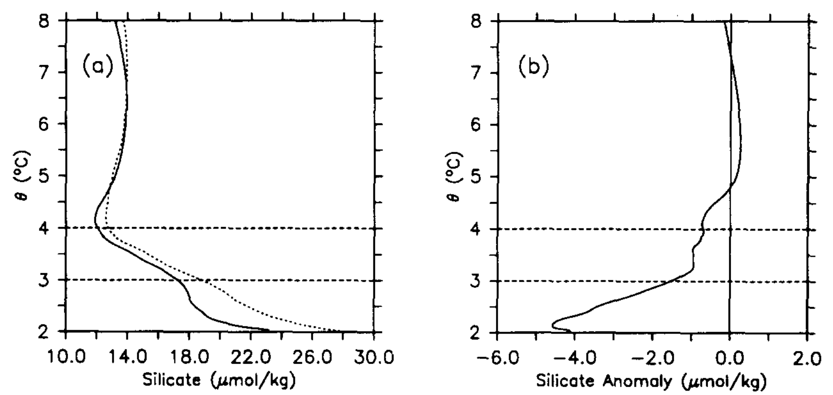

Fig. 6. (a) Average profile of silicate vs potential temperature (colder than 8°C) for all the stations in the DWBC (see Fig. 1). (b) Silicate anomaly.

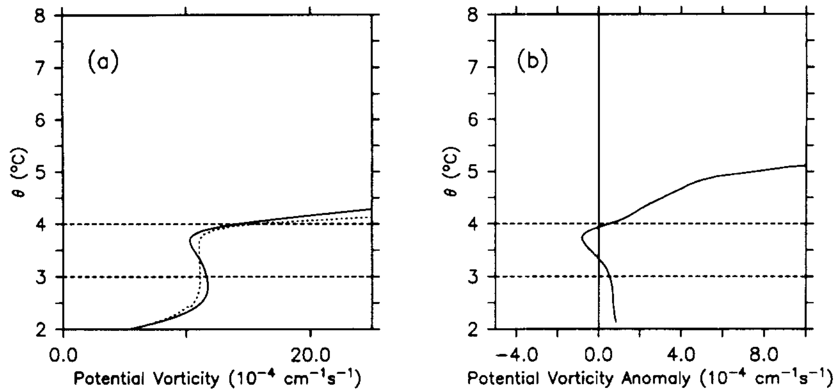

Fig. 7. (a) Average profile of potential vorticity vs potential temperature (colder than 8°C) for all the stations in the DWBC (see Fig. 1). (b) Potential vorticity anomaly.

the Labrador Sea (Table 1) are compared with the concentrations observed west of the Grand Banks in OC134, the per cent concentration change over this distance (roughly 2500 km) is very large for silicate and F-11 but small for salinity and oxygen. The question then arises, are these differences consistent with regard to the dilution of the various tracers? This can be answered with the following simple calculation.

Consider a steady state tracer balance in the DWBC between advection and lateral mixing,

$$u\theta_x = \kappa\theta_{yy}, \tag{1}$$

where x is alongstream and y is cross-stream, u is the magnitude of the current, κ is the constant lateral diffusivity, and θ is the tracer concentration. We crudely approximate the gradients in equation (1) by $\theta_x \sim (\theta_d - \theta_u)/\Delta x$, and $\theta_{yy} \sim (\theta_i - \theta_u)/\Delta y^2$, where θ_u, θ_d, θ_i are the upstream, downstream, and interior tracer concentractions. Then equation (1) can be rewritten,

$$\left(\frac{\theta_d - \theta_u}{\theta_i - \theta_u}\right) \sim Pe^{-1}, \tag{2}$$

Table 1. Downstream evolution of four tracers in the deepest component of the DWBC. Column 1 lists the core values at 54°N (Wallace and Lazier, 1988), and column 2 lists the core values from the OC134 composite profiles (centered at 64°W). Column 3 contains the interior values from the OC134 interior profiles (i.e. at the appropriate value of θ). Column 4 lists the per cent downstream change for each property, and column 5 lists the inverse Peclet number

Property	θ_u	θ_d	θ_i	$\left\lvert\frac{\theta_d - \theta_u}{\theta_u}\right\rvert$	$\left(\frac{\theta_d - \theta_u}{\theta_i - \theta_u}\right) = Pe^{-1}$
Salinity	34.90	34.923	34.925	0.001	0.9
Oxygen	6.7	6.36	6.26	0.05	0.8
Silicate	10	19.6	24.0	0.98	0.7
F-11	2.5	0.35	0.12	0.86	0.9

where $Pe = u\Delta y^2/\Delta x\kappa$ is a Peclet number which measures the relative strength of advection vs diffusion. Note that $0 \leq Pe^{-1} \leq 1$; the lower bound $Pe^{-1} = 0$ represents the advective limit ($\theta_d = \theta_u$, i.e. no tracer diffuses into the interior) and the upper bound $Pe^{-1} = 1$ is the diffusive limit ($\theta_d = \theta_i$, i.e. the maximum amount of tracer diffuses into the interior). Although the per cent change in concentration varies drastically as noted, the value of Pe^{-1} is nearly the same for each tracer (near the diffusive limit, Table 1), indicating that the observed dilution is indeed consistent among the various tracers. The reason for the large observed downstream change in F-11 vs oxygen for instance is simply because the interior water is nearly void of F-11 but contains a fairly high concentration of oxygen.

Classical Labrador Sea water

Above the deep overflow water in the DWBC is the water mass known as Labrador Sea water (LSW), which is formed convectively during winters in the central Labrador Sea. This water spreads into different regions of the North Atlantic (TALLEY and MCCARTNEY, 1982), part of it progressing equatorward within the DWBC. LSW is traceable away from its source by its fresh salinity and low potential vorticity (indicative of convective formation) (WARREN and VOLKMANN, 1968; TALLEY and MCCARTNEY, 1982). At mid-latitudes its core temperature is approx. 3.5°C (TALLEY and MCCARTNEY, 1982) and core density $\sigma_{1.5} \sim 34.67$ kg m^{-3} (all densities referred to here are 1980 equation of state).

The formation of LSW is closely tied to the various water masses in the Labrador Sea and their respective circulations (Fig. 8). The West Greenland current transports near-surface fresh water (which originally came from the East Greenland current); below the surface, however, the West Greenland current has a warm, salty core due to entrainment of nearby Irminger Sea water. Part of the West Greenland current turns westward, flows across the northern end of the Labrador Sea and then combines with extremely fresh water of polar origin to form the Labrador current. To the south the Labrador Sea is bounded by the North Atlantic current (the northern extension of the Gulf Stream). In the deep layer the NGOW progresses cyclonically around the Labrador Sea.

The surface waters of the central Labrador Sea are continually kept fresh by spreading from the Labrador current. The circulation in the central part of the basin consists of a cyclonic gyre (LAZIER, 1973) which is strengthened during winter, causing the isotherms to rise and reducing the stratification (CLARKE and GASCARD, 1983). A large heat loss to the atmosphere then drives deep convection which mixes the fresh surface water downward and tags the LSW with its low salinity and low potential vorticity.

It is well known that the formation of LSW occurs on a sporadic basis. Ocean weather ship (O.W.S.) *Bravo* was maintained in the central Labrador Sea from 1945 to 1974, providing a time series of oceanic and atmospheric conditions in the central basin. Using these data TALLEY and MCCARTNEY (1982) showed that there was a 10-year period from 1962 to 1971 during which deep convection was diminished and as a result LSW was isolated from renewal. LAZIER (1980) also analysed the *Bravo* data set for roughly the same 10-year period and showed that during this time the depth to which convection penetrated was highly variable, from 200 to 1500 m. Although the *Bravo* ocean station is no longer occupied, there have been more recent observations of deep convection in the central basin (e.g. CLARKE and GASCARD, 1983; WALLACE and LAZIER, 1988).

Consistent with what TALLEY and MCCARTNEY (1982) saw farther to the north, there is a relative minimum of potential vorticity $(f/\rho)(\delta\rho/\delta z)$ in the 3–4°C temperature range in the

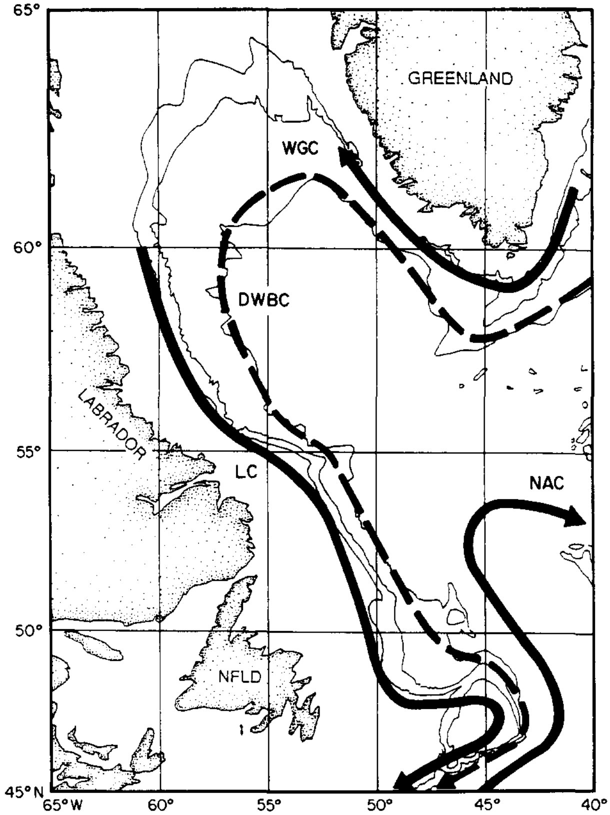

Fig. 8. Schematic representation of the major currents in the region of the Labrador Sea: WGC, West Greenland current; LC, Labrador current; NAC, North Atlantic current; DWBC, deep western boundary current.

OC134 data (Fig. 7a). This feature of the mid-depth component of the DWBC is boundary intensified as evidenced by the anomaly profile of Fig. 7b. Note that although the deeper NGOW has low potential vorticity as well, its values are actually high relative to the interior (Fig. 7b), a fact previously noted by Hogg *et al.* (1986). In addition to its low potential vorticity, LSW also has been identified at mid-latitudes by its high oxygen content (e.g. Clarke *et al.*, 1980; Richardson, 1985), which is indicative of recent contact with the atmosphere. However, oxygen is not a distinguishing tracer of the LSW far from the source region (Fig. 3a). This is also true of F-12, which has a relatively low value within the LSW (Fig. 4a). This is not to say that LSW does not contain high concentrations of these properties, indeed newly formed LSW is characterized by a distinct CFC signal (Wallace and Lazier, 1988). However, farther to the south the LSW signal is overwhelmed by the large CFC extrema of the NGOW and shallow DWBC component.

4–5°C *water*

The shallowest component of the DWBC is a water mass characterized most readily by a core of high CFCs and tritium in the 4–5°C temperature range. In contrast to the two deeper components of the DWBC, many basic questions still exist regarding this water mass, including its mean equatorward transport and precise origin. This water has been identified as modified LSW (WEISS *et al.*, 1985; FINE and MOLINARI, 1988); however, we show below that it is not dense enough to be formed in the central Labrador Sea even during anomalous conditions. Instead, based on density and property characteristics, the area of formation appears to be the southern Labrador Sea inshore of the North Atlantic current.

The CFC profile (Fig. 4a) suggests that the 4–5°C water is ventilated more consistently than both the deeper LSW component and the water directly above. WORTHINGTON (1976) defined a fresh water mass progressing equatorward against the western boundary with core temperature 6°C as Subarctic Intermediate water (SAIW). The high CFC shallow DWBC water is distinctly colder than this; however, one might be tempted to call the 6–8°C water in our data set (with lower CFC concentration) SAIW, especially since it is somewhat fresh, as will be seen below. However, WORTHINGTON (1976) maintained that SAIW usually did not penetrate west of the Grand Banks (although there are occasional exceptions, e.g. FUGLISTER, 1963). Thus, we will refrain from using the term SAIW except when referring to northerly regions where this water mass is clearly identifiable.

Interestingly, the 4–5°C DWBC water does not have an associated high oxygen signal (Fig. 3a), whereas the NGOW has both a high CFC core and high oxygen core. This is most likely because the 4–5°C water is in close proximity to the pronounced oxygen minimum layer which exists due to biological consumption (WYRTKI, 1962). It should be noted that the slight oxygen maximum at 3.8°C (Fig. 3a) is a real feature readily apparent in the raw data. Because this maximum coincides precisely with the potential vorticity minimum of the LSW (Fig. 7), one is tempted to correlate these features as LSW core properties. However, an alternate explanation is that the increase in oxygen at temperatures warmer than 3.5°C gets truncated by mixing with the above-lying low oxygen water. This view is more consistent with the F-12 profile which in the absence of consumption continues to increase at even warmer temperatures. The oxygen anomaly profile (Fig. 3b) also supports this latter view; it clearly shows that relative to the interior, LSW has a relative oxygen minimum like that of F-12.

Because of the absence of an oxygen signal, the anomalous nature of the warmest DWBC component was not recognized until the anthropogenic tracers, CFCs (and tritium), became available. It turns out, however, that at mid-latitudes this water also can be identified by its anomalously fresh salinity which distinguishes it from the waters above and below.

T–S properties. The F-12 maximum in the shallow component of the DWBC is centered at $\theta = 4.7$°C, $\sigma_\theta = 27.68$ kg m^{-3} and $\sigma_{1.5} = 34.52$ kg m^{-3} (Fig. 9). By contrast, classical LSW has its core near $\theta = 3.5°$, $\sigma_{1.5} = 34.67$ kg m^{-3} (TALLEY and MCCARTNEY, 1982), clearly denser than this F-12 core. Tritium measurements collected along the continental boundary also reveal the high tritium signal of the same 4–5°C water (e.g. JENKINS and RHINES, 1980; OLSON *et al.*, 1986).

The high CFC–tritium core also coincides with a core of anomalously low salinity at sections 1, 3 and 4 of OC134 (Fig. 10). At section 2, however, there is an exceptionally large fresh anomaly near 7°C (400 m) that is not present at the other sections. The anomalous nature of this feature is clearly seen in the salinity vs depth profiles (Fig. 11), and it may be in fact a patch of true SAIW. Similar patches have been observed in this area at similar densities (McCartney *et al.*, 1980), and this is consistent with Worthington's (1976) notion that SAIW does not regularly penetrate west of the Grand Banks. If and when it does travel this far west, however, its pronounced salinity anomaly apparently overwhelms the deeper anomaly of the 4–5°C water. Note that the F-12 concentration in the SAIW patch (Fig. 10, section 2) is anomalously high compared with the other three profiles at this depth. This suggests that such patches of SAIW contain more recently ventilated water.

Tracing the 4–5°C DWBC component by its low salinity is not always straightforward. The bounding water masses are both anomalously fresh as well (negative values in Fig. 10), so the division between water masses is not always identifiable. This, together with the fact that there is no oxygen signal, is probably why the 4–5°C DWBC component was not previously singled out in the literature. Nonetheless, close inspection of historical hydrographic sections at mid-latitudes often reveals the anomalous salinity of this component against the continental slope. Clarke *et al.* (1980) described a region of low salinity at 50°W in the σ_θ range of the shallow DWBC (Fig. 12), but they did not have CFC or tritium data that might have suggested that the feature be distinguished from the deeper LSW. Extremely fresh SAIW is also identifiable in Fig. 12 as the fresh tongue shallower than 500 m near Sta. 61 and Sta. 62 (Clarke *et al.*, 1980).

South of about 30°N the ability to track the shallow DWBC component by its salinity signal becomes difficult. This water mass encounters both the anomalously fresh Antarctic Intermediate water (Fine and Molinari, 1988) and salty Mediterranean water (Weiss *et al.*, 1985), and by the time the shallow DWBC reaches the equator the high CFC core coincides with a maximum in salinity (Weiss *et al.*, 1985). Fine and Molinari (1988) illustrate how the presence of Antarctic Intermediate water causes a mid-depth salinity maximum to be formed (their fig. 7).

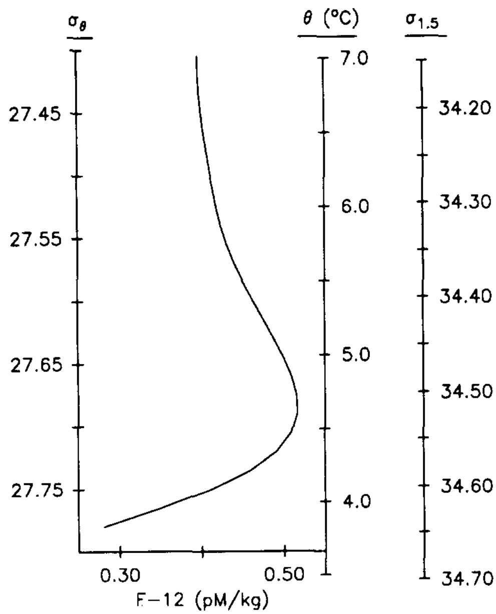

Fig. 9. The CFC maximum of the shallow DWBC vs potential temperature and density.

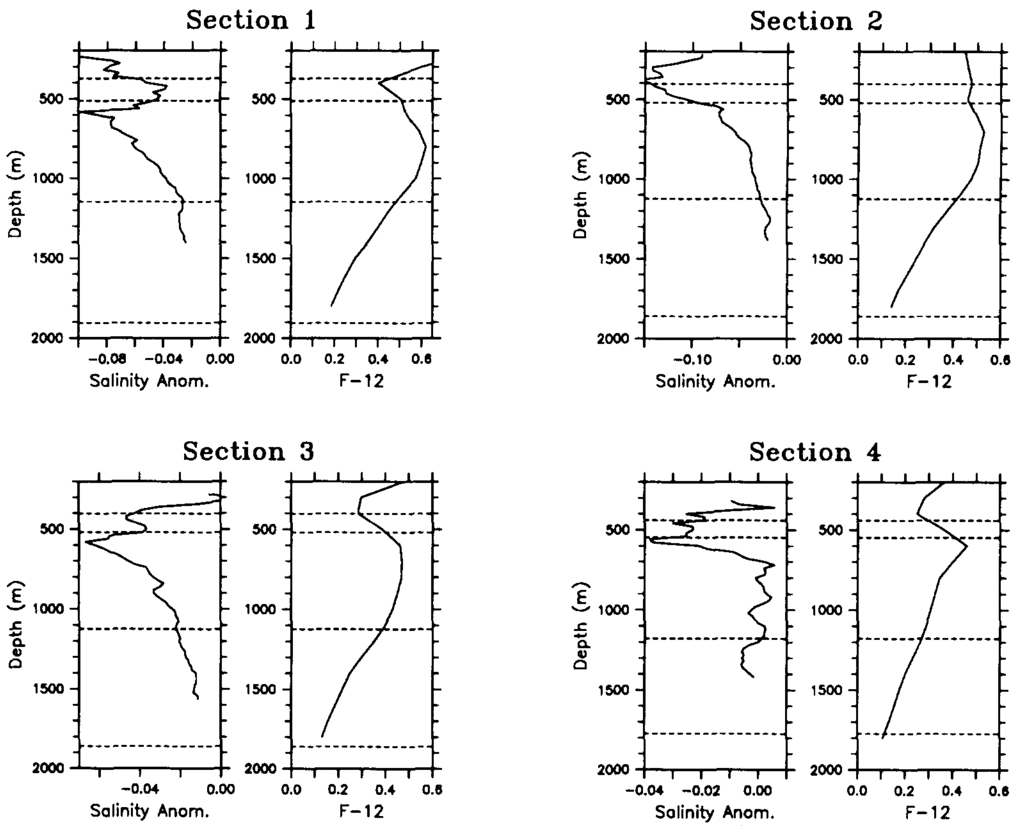

Fig. 10. Average profiles of salinity anomaly (left) and F-12 (right) vs depth for the individual sections of OC134. The lower set of dashed lines marks the potential temperature range 3.5–4°C and the upper set marks the 5.5–7°C temperature range.

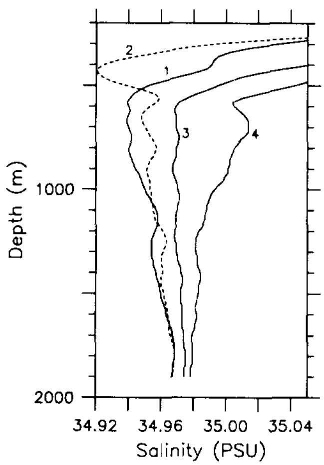

Fig. 11. Average profile of salinity vs depth for each section. Note the anomalously fresh feature near 400–500 m at section 2 (dashed line).

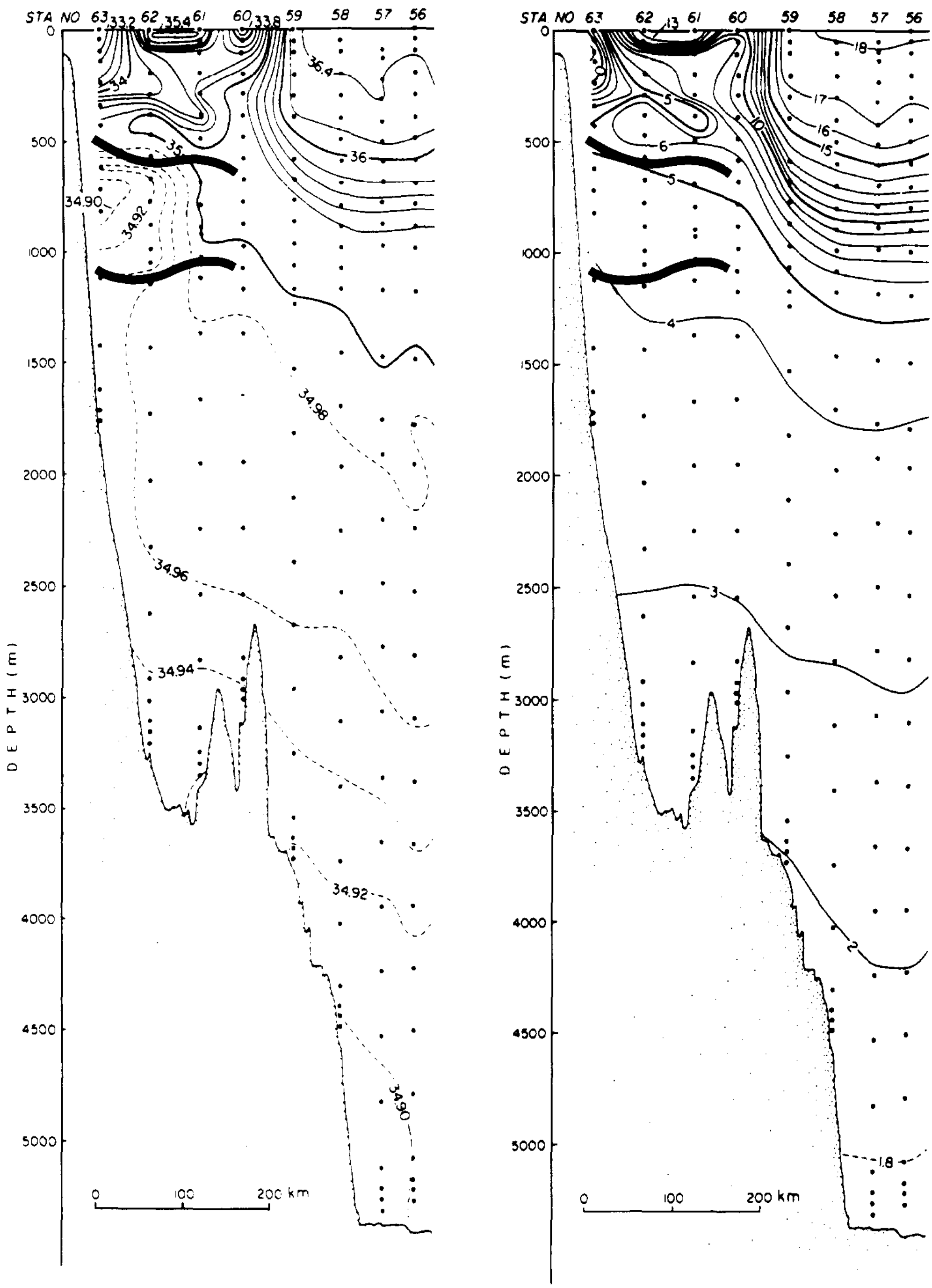

Fig. 12. Vertical section of salinity and potential temperature at 50°W (from CLARKE *et al.*, 1980). The density interval corresponding to the shallow CFC core ($27.60 \le \sigma_\theta \le 27.73$ kg m^{-3}) is delineated by thick lines near the western boundary.

Circulation. The high CFC–tritium signal of the 4–5°C water has been identified against the continental slope throughout the North Atlantic (e.g. WEISS *et al.*, 1985; FINE and MOLINARI, 1988; SMETHIE, submitted). The mean equatorward transport and flow speed of this component of the DWBC remain undetermined, however. Using a CFC ratio argument WEISS *et al.* (1985) estimated its mean flow to be only 1 cm s^{-1}, and a similar estimate is obtained from the OC134 data set (SMETHIE, submitted). By contrast, FINE and MOLINARI (1988) found that the high CFC core roughly coincided with a mid-depth geostrophic jet on the order of 20 cm s^{-1}. RICHARDSON (1977) also measured a subsurface boundary current of roughly 20 cm s^{-1} in the 4–5°C temperature range coincident with a core of low salinity anomaly.

Using historical current meter data from 40 locations along the mid-Atlantic Bight, WATTS (1991) computed an average velocity section of the DWBC. The section shows the shallowest component of the DWBC flowing equatorward at 5–10 cm s^{-1}, compared with the LSW at 2–5 cm s^{-1}, and the NGOW at 5–7 cm s^{-1}. WATTS (1991) also computed the mean equatorward transports for the different DWBC components, although there is more uncertainty in these values because of the subjectiveness in choosing the offshore limit of the current. WATTS (1991) quotes 2–6 Sv for the 4–6°C temperature range, 1–3 Sv for the LSW and 1.7–5 Sv for water deeper than 2500 m.

ORIGIN OF THE 4–5°C COMPONENT

The shallow component of the DWBC has its origin at northerly latitudes (e.g. JENKINS and RHINES, 1980; WEISS *et al.*, 1985; FINE and MOLINARI, 1988). WEISS *et al.* (1985) using an isopycnal analysis claimed that it was slightly warm LSW; however, they chose a density surface ($\sigma_{1.5}$ = 34.63 kg m^{-3}) that is clearly too dense (see Fig. 9). FINE and MOLINARI (1988) recognised this and instead referred to the water as incompletely formed LSW. They claimed that it was formed sometime during the warm 10-year period in the central Labrador Sea documented by O.W.S. *Bravo*. This is also unlikely, as we now explain.

The concept of incompletely formed LSW was put forward by TALLEY and MCCARTNEY (1982) to explain observations of particularly warm temperatures associated with the low potential vorticity signal of the LSW. They explained that less intense convection in the central Labrador Sea during the warm period from 1962–1971 produced a lighter variety of LSW. However, this water cannot be the source of the high CFC–tritium DWBC water because the upper waters in the central Labrador Sea are simply too dense. In seven out of the 10 winters during the warm period the water as shallow as 10 m depth at ocean station *Bravo* was denser than σ_θ = 27.68 kg m^{-3} (i.e. the core density of the high CFC–tritium water, Fig. 9) (LAZIER, 1980). Thus, only during the three warmest winters in a three-decade span was there a small amount of surface water of sufficiently light density, and there are historical sections revealing the low salinity 4–5°C water at mid-latitudes prior to these warm winters (e.g. WARREN and VOLKMANN, 1968).

The potential vorticity minimum in the OC134 data (Fig. 7) occurs at 3.8°C, $\sigma_{1.5}$ = 34.65 kg m^{-3}, which is lighter than the classical LSW. This is most likely the incompletely formed LSW having penetrated west of the Grand Banks. Because the OC134 data were collected more than a decade after the warming trend, a very small advective speed for the LSW component of the DWBC (0.5 cm s^{-1}) is implied. However, the process by which newly formed LSW progresses from the central Labrador Sea into the DWBC is not understood, and the new water may reside in the basin for a while before progressing equatorward (perhaps recirculating in the cyclonic gyre). Another possibility is that the LSW observed in 1983 was not formed during the 10-year warm trend but was formed more recently during an isolated occurrence of incomplete convection. Because of the lack of data coverage in the Labrador Sea after O.W.S. *Bravo* this possibility cannot be ruled out.

The continuity of the 4–5°C CFC–tritium signal in the North Atlantic indicates that this water mass is being formed on a regular basis. Since it is not formed in the central Labrador Sea, either by complete or incomplete convection, where does it come from? According to McCartney and Talley (1982), subpolar mode waters recirculate in a cyclonic path around the northern North Atlantic, becoming successively more dense due to air–sea buoyancy exchange, and finally reaching the density of LSW during deep convection in the central Labrador Sea. The particular class of mode water with density comparable to the high CFC–tritium DWBC water is formed in the Irminger Sea east of Greenland. This water is high in salinity, and it is most likely not the source of the shallow DWBC for several reasons.

Firstly, in order for this water to reach mid-latitudes in large quantities without being converted to denser water, deep convection must abate in the Labrador Sea; only then might the unmodified Irminger water flow around the perimeter of the Labrador Sea and equatorward. In light of the historical data, however, the absence of deep convection in the Labrador Sea appears to be the exception not the rule. Furthermore, data collected in 1966 (see Grant, 1968) in the middle of the warm period show very little warm Irminger water flowing out of the Labrador Sea. Secondly, the shallow DWBC becomes more saline as it progresses equatorward (Fig. 11) presumably from mixing with the very saline North Atlantic current and Gulf Stream; the Irminger mode water is salty to begin with and would only become more saline. Finally, lack of any atmospheric contact in the Labrador Sea would mean a longer period of isolation of the mode water, thus any recently renewed denser LSW would have a relatively higher CFC content. Despite the fact that dense LSW has been formed in the 1970s and 1980s (e.g. Clarke and Gascard, 1983; Wallace and Lazier, 1988) its CFC concentration remains lower than the warmer 4–5°C water.

Clarke and Gascard (1983) noted the importance of the salty entrained Irminger Sea water in the West Greenland current in maintaining a decreased level of stratification in the upper waters of the central Labrador Sea, which facilitates the onset of deep convection. Lazier (1980) showed that during the three particularly anomalous winters (1968–1970) of the 10-year warm period the surface waters at O.W.S. *Bravo* were markedly more fresh than in the remaining seven winters. He argued that it was this fresh surface layer that increased the near-surface stratification and helped limit the convection. This implies that the area of formation of the shallowest component of the DWBC should be characterized by particularly low near-surface salinities. This would permit convection only to shallower depths, and the water so formed would be imparted with a particularly low salinity anomaly. This is consistent with the shallow DWBC water observed to the south.

A lateral distribution of salinity at 50 m in the Labrador Sea (Fig. 13a) shows that the lowest interior salinities occur in the southern Labrador Sea (<34.8 PSU), inshore of the North Atlantic current. As with the central Labrador Sea, the source of this low salinity water to the south is the Labrador current. The reason why the water is fresher to the south is that there is less influence from the saline West Greenland current (which borders the northern Labrador Sea both to the east and north). Figure 13b shows where the

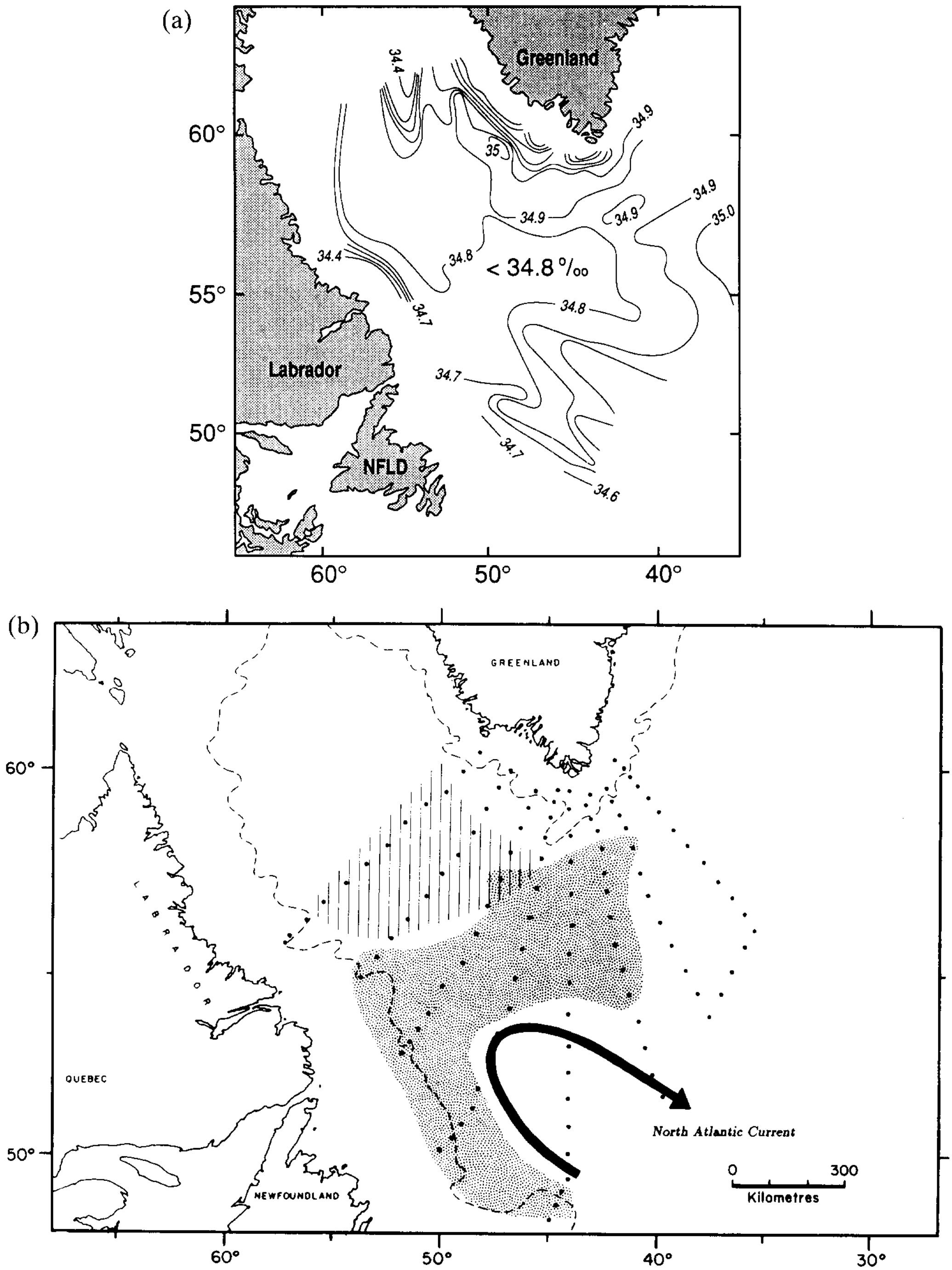

Fig. 13. (a) Distribution of salinity at 50 m depth in the Labrador Sea (from LAZIER, 1973). (b) Region of the Labrador Sea (indicated by shading) where the temperature is warmer than 3.8°C at 400 m and colder than 5.5°C within the upper 400 m (which serves to exclude the water of the North Atlantic current). Stations where the salinity is greater than 34.9 PSU within the upper 400 m (the West Greenland current) or less than 34.7 PSU (the Labrador current) are not included. The cross-hatched region indicates σ_t at 50 m > 27.68 kg m^{-3} (the core density of the shallow DWBC).

temperature at 400 m in the interior is warmer than the upper limit of newly formed LSW; this is roughly the same area as in Fig. 13a. The part of the central Labrador Sea where upper layer densities are too great to form the shallow DWBC water is marked as well in Fig. 13b. Eastward of 40°W temperatures are too warm, salinities too high (Fig. 13a) and densities too light to be the formation area. Thus, based on T–S characteristics and density, the formation region of the 4–5°C DWBC component is most likely the region of the southern Labrador Sea inshore of the North Atlantic current, west of 40°W.

It should be noted that the data used to make Fig. 13a and b come from a single survey conducted in 1966 (see GRANT, 1968), and in light of the observed interannual variability that occurs in the Labrador Sea this single realization should not be taken as representative. The winter of 1966 was in the middle of the 10-year warm trend and near-surface salinities throughout the Labrador Sea were fresher than in previous years. However, the important feature is the presence of the relative minimum in salinity to the south, a feature that also was present in 1962 just prior to the warming trend (WORTHINGTON and WRIGHT, 1970). A hydrographic section extending southward from the tip of Greenland occupied in April 1978 (CLARKE, 1984) revealed that the low salinity water inshore of the North Atlantic current (in the right density range) was fresher and colder than in 1966. At mid-latitudes, however, the characteristics of the shallow DWBC water mass show little variation; for instance, the potential temperature of the tritium maximum observed in 1977 (JENKINS and RHINES, 1980) is the same as in the OC134 data (1983), and also the same as that observed in 1986 (SMETHIE, submitted). Whether this means that mixing and entrainment after formation serve to keep downstream properties somewhat constant or that the downstream sampling is too sparse remains an open question.

Downstream mixing

The large difference in salinity between the very fresh surface water in the southern Labrador Sea (Fig. 13a) and the shallow DWBC water observed west of the Grand Banks in OC134 (Fig. 11) implies that substantial mixing occurs subsequent to the formation of this water. As the DWBC progresses southward from the Labrador Sea then around the Grand Banks it is in close proximity to the saline North Atlantic current (and Gulf Stream further west). It has been shown that intense mixing occurs across the front of the North Atlantic current (GEORGI and SCHMITT, 1983) and that such mixing can account for the observed downstream property changes of the North Atlantic current (CLARKE *et al.*, 1980; GEORGI and SCHMITT, 1983). The converse of this is that the southward progressing northern waters will have their properties altered as well. The following simple calculation gives an idea as to the mixing required to modify the salinity of the northern source water to that which is observed at 50°W (Fig. 12) versus the advective strength of the shallow DWBC component.

If one assumes a steady balance between the southward advection and lateral mixing then equation (1) applies, which can be approximated as

$$u\Delta S_x/\Delta x \sim \kappa \Delta S_y/\Delta y^2, \tag{3}$$

where ΔS_x is the alongstream change in salinity and ΔS_y is the cross-stream change. Both u, the shallow DWBC flow speed, and κ, the constant lateral mixing coefficient, are unknown. We take $\Delta y \sim 100$ km (Fig. 12) and $\Delta x \sim 1200$ km (the distance from the southern Labrador Sea to 50°W). For the initial salinity we use 34.82 PSU, which is a representative value within the upper 400 m in the southern Labrador Sea; the downstream value is 34.90 PSU (Fig. 12). To compute the average cross-stream salinity gradient we use 35.00 PSU for the offshore value (Fig. 12) and use the average of the upstream and downstream values for an onshore value. Putting these numbers into equation (3) results in the following relation for u and κ,

$$u = (2.1 \times 10^{-6}\,\mathrm{cm}^{-1})\kappa. \tag{4}$$

Using microstructure observations from the North Atlantic current north of the Grand Banks, GEORGI and SCHMITT (1983) calculated a value of κ of the order of 10^7 cm^2 s^{-1}, which is also what CLARKE *et al.* (1980) used in their analysis. With such a κ, equation (4) gives $u = 20$ cm s^{-1}, which is reminiscent of the above mentioned hydrographic estimates of the shallow DWBC core speed. For $\kappa = 10^6$ cm^2 s^{-1}, u decreases to 2 cm s^{-1} which is more like the CFC ratio calculation noted above. These core speeds probably represent the upper and lower extremes and show that the observed alongstream change in salinity is consistent with a southern Labrador Sea source and consistent as well with what little we know about the flow strength and mixing of the shallow DWBC.

As mentioned in the previous section, it is puzzling that downstream from the source region the shallow DWBC contains high concentrations of CFCs and tritium, but low oxygen concentration. Since the interior water contains comparable or higher oxygen at these temperatures (Fig. 3b), lateral mixing cannot account for the large reduction in oxygen from the source (it is presumed that at formation the shallow DWBC water is oxygen-rich). The decrease in oxygen most likely occurs because the shallow DWBC resides just below the oxygen minimum layer.

To investigate this consider a balance between advection, vertical diffusion and consumption,

$$u\Delta O2_x/\Delta x \sim \nu\Delta O2_z/\Delta z^2 - R, \tag{5}$$

where $\Delta O2_x$, $\Delta O2_z$ are the alongstream, vertical changes in oxygen, ν is the constant vertical diffusivity and R is the biological consumption. For the source concentration we use the oxygen saturation value of 7.3 ml l^{-1} for the appropriate temperature and salinity (3.6°C, 34.82 PSU). The downstream value is 6.4 ml l^{-1} (CLARKE *et al.*, 1980), and $\Delta O2_z = -2.9$ ml l^{-1} for $\Delta z = 300$ m (CLARKE *et al.*, 1980). For R we use the value of 0.032 ml l^{-1} y^{-1} calculated by JENKINS (1980) for the appropriate density layer. The value of the vertical diffusivity for which mixing is as important as consumption is $\nu = 0.3$ cm^2 s^{-1}. For diffusivities much less than this the two-term balance of advection and direct consumption gives a DWBC core speed of only 0.15 cm s^{-1}, which is an order of magnitude smaller than the lower bound obtained from the above salinity balance. For a vertical diffusivity of $\nu = 1$ cm^2 s^{-1} the mixing dominates consumption and the resulting core speed is 0.5 cm s^{-1}. There is evidence that ν may be even larger than this in the DWBC (PICKART and HOGG, 1989), and the oxygen concentraction may not be saturated at formation; both of these would lead to a larger estimate of the core speed. It is evident that the precise evolution of the shallow DWBC oxygen signal needs to be addressed more carefully with appropriate data near the source. The results here do indicate, however, that it is vertical mixing with oxygen depleted water from above rather than direct consumption which lowers the shallow DWBC oxygen signal.

SUMMARY

Using hydrographic data collected at mid-latitudes, the three components of the DWBC—the NGOW, LSW and 4–5°C water—were resolved and their distinguishing characteristics analysed. For the LSW the only truly distinguishing tracer is the potential vorticity, as the other historical tracers of this water mass, oxygen and salinity, do not uniquely define it at mid-latitudes. The other two DWBC components are revealed by CFCs and tritium, which implies that these water masses have been ventilated more recently than the LSW which lies between them. The NGOW also can be traced by its high oxygen and low silicate anomaly, whereas the shallowest DWBC component can be identified by its anomalously low salinity.

The source locations and formation of the two deeper DWBC components are fairly well understood. The LSW is formed by deep convection in the central Labrador Sea, and the NGOW enters the basin via the northern sills to form the deepest core. In contrast to earlier descriptions, we have shown that the shallowest component of the DWBC is not a modified variety of LSW; it is not dense enough to be formed by convection in the central Labrador Sea. Rather, its T–S character and tracer content suggest that it is formed in a region of the southern Labrador Sea inshore of the North Atlantic current which is characterized by particularly low upper layer salinities. After formation it progresses equatorward adjacent to the saline North Atlantic current; the intense lateral mixing which occurs in this region is sufficient to modify the salinity of the shallow DWBC core to that which is observed at mid-latitudes.

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