Labrador Sea Boundary Currents and the Fate of the Irminger Sea Water

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ABSTRACT

The general circulation of the Labrador Sea is studied with a dataset of 53 surface drifters drogued at 15 m and several hydrographic sections done in May 1997. Surface drifters indicate three distinct speed regimes: fast boundary currents, a slower crossover from Greenland to Labrador, and a slow, eddy-dominated flow in the basin interior. Mean Eulerian velocity maps show several recirculation cells located offshore of the main currents, in addition to the cyclonic circulation of the Labrador Sea. Above the northern slope of the basin, the surface drifters have two preferential paths: one between the 1000-m and 2000-m isobaths and the other close to the 3000-m isobath. The vertical shear estimated from CTD data supports the presence of two distinct currents around the basin. One current, more baroclinic, flows between the 1000-m and 2000-m isobaths. The other one, more barotropic, flows above the lower continental slope. The Irminger Sea Water carried by the boundary currents is altered as it travels around the basin. Profiling Autonomous Lagrangian Circulation Explorer (PALACE) floats that followed approximately the Irminger Sea Water in the Labrador Sea show signs of isopycnal mixing between the interior and the boundary current in summer–fall and convection across the path of the Irminger Sea Water in winter–spring.

1. Introduction

The shallow water boundary circulation of the Labrador Sea is essentially cyclonic and consists of three main currents (Fig. 1). The West Greenland Current flows northward along the Greenland shelf and shelf break with a transport of roughly 3 Sv (Sv = 10⁶ m³ s⁻¹) measured at Cape Farewell (Clarke 1984). Along the Labrador coast, the shallow Labrador Current flows southward above the shelf and upper slope with a summer transport around 11 Sv and an annual range of 4 Sv (Lazier and Wright 1993). Most of the transport occurs in a jet centered at the shelf break whose velocity is maximum in October and minimum in March–April due to variations in freshwater influx from Baffin Bay (Lazier and Wright 1993). The “Northwest Corner” of the North Atlantic Current limits the southern boundary of the Labrador Sea. The West Greenland Current carries fresh and cold water (θ ~ −1.8°C, S ≤ 34.5 psu) from the Nordic seas. The warmer and saltier Irminger Sea Water (ISW: θ ~ 4.5°C, S ~ 34.95 psu) flows above the upper slope and below the cold water. The ISW flow has been estimated at 11 Sv by Clarke (1984). Similarly the Labrador Current carries fresh and cold water from Baffin Bay (θ ~ −1.5°C, S ≤ 34 psu) above the shelf-break area. Modified ISW that has traveled around the basin is found below the Baffin Bay water, along the upper Labrador slope. Offshore of the main boundary currents, a series of cyclonic recirculation cells form a weak anticyclonic flow in the interior (Lavender et al. 2000). The deep circulation consists of the deep western boundary current traveling cyclonically along the 3000-m isobath. It carries North Atlantic Deep Water, including northeast Atlantic deep water (θ ~ 3°C, S ~ 34.92 psu) and the denser Denmark Strait overflow water (θ ≤ 1.5°C, S ~ 34.9 psu). The total transport past Cape Farewell ranges between 34 and 50 Sv (Clarke 1984; Gana and Provost 1993; Reynaud et al. 1995).

In addition to the “classical” Labrador Current associated with the shelfbreak density front centered at the 1000-m isobath, Lazier and Wright (1993) observed
a current with a significant barotropic component near the 2500-m isobath, which they named the “deep” Labrador Current. They suggest that this current is part of the large-scale subpolar gyre circulation that has to be offshore of the shelfbreak Labrador Current (Thompson et al. 1986). Lazier and Wright (1993) observed that the deep Labrador Current is stronger in winter and weaker in summer, which agrees with the results of Greatbach and Goulding (1989) who obtained an enhanced subpolar gyre in January–February and a weaker gyre in July from a barotropic linear model of the North Atlantic forced with climatological wind stress. However, Böning et al. (1996) argue from the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM) experiments in favor of a significant overflow-driven component of the subpolar gyre.

The Labrador Sea plays a significant role in the thermohaline circulation as the formation site of the Labrador Sea Water (LSW). When the strong winter northwesterly winds from the Canadian Arctic reach the Labrador Sea ice edge, they generate high air–sea heat fluxes that erode the surface stratification and allow deep vertical overturning (Marshall and Schott 1999) Clarke and Gascard (1983) suggested that the deepest convection takes place north of Hamilton Bank and close to the Labrador slope. The ISW is essential to the convection process as it is the heat source balancing the wintertime cooling to the atmosphere. It is thus a part of the restratification process after convection and it keeps the Labrador Sea ice free (Lilly et al. 1999; Lazier 1973).

The ISW is a remnant of the subpolar mode water (McCartney 1992; McCartney and Talley 1982, 1984), the warm water of the Gulf Stream that has traveled cyclonically around the subpolar gyre. McCartney (1992) suggests that about 15 Sv of subpolar mode water are converted to intermediate or deep water around the subpolar gyre. The “new” Labrador Sea Water formed in winter by deep convection is the final conversion product of the subpolar mode water.

Most observational and modeling studies on the Labrador Sea have been focused on the Labrador Current, with less known about the other components of the boundary circulation. Using mixed layer drifters drogued at 15 m released during the 1990s in the North Atlantic subpolar gyre and several hydrographic sections, we have studied the vertical and horizontal structure of the boundary currents (sections 3 and 4). Similarly to Lavender et al. (2000), we observed recirculation cells offshore of the main boundary current system. Two PALACE floats entrained in the boundary currents were used to track the ISW around the basin (section 5a). The evolution through the 1990s of the Irminger Sea Water potential temperature and salinity at the Greenland and Labrador slopes is described in section 5b. The datasets are described in section 2, and a summary is included in section 6.

2. Data description

a. Surface drifters

The surface drifters are of the WOCE/TOGA (World Ocean Circulation Experiment/Tropical Ocean Global Atmosphere) type and were released during the 1990s over the North Atlantic. The drifters consist of a surface float attached to a holey-sock drogue centered at 15 m. With a drag area ratio of about 39, these floats are expected to have a downwind slip of less than 0.1% of the wind speed for magnitudes under 10 m s$^{-1}$ (Niiler et al. 1995). However, the lack of data for wind exceeding 10 m s$^{-1}$ prevents any conclusion concerning weak currents under stronger winds. The drifters were tracked by satellite with the ARGOS system with a precision of 300 m. The data were despiked, interpolated onto 6-h intervals (Hansen and Poulain 1996), and low-pass filtered in order to remove the tidal and inertial components. The wind-slip component of the velocity was estimated and removed from the data following...
Fig. 2. Trajectories of surface drifters, drogued at 15 m, in the North Atlantic subpolar gyre. Circles and triangles mark, respectively, the first and the last recorded fixes for each drifter. Each drifter is assigned a color with respect to its launching region (Gulf Stream and Newfoundland Basin: yellow; European Basin: green; Iceland and Irminger Basin: magenta; Labrador Basin: blue). The four regions defined are delimited by the thick black dashed lines.
Fig. 3. Drifters that went through the Labrador Sea region. Transit times and speeds given in Table 1 were estimated between the thick black lines (Segments 1 to 4 going counterclockwise).

Niiler and Paduan (1995): $U_{\text{slip}} = (0.07/R)W$, $U_{\text{slip}}$ being the wind-slip component of the velocity in the direction of the wind, $R$ the drag area ratio, and $W$ the wind speed.

b. PALACE floats

As a U.K. contribution to WOCE, six PALACE (Profiling Autonomous Lagrangian Circulation Explorer) floats were released in October 1996 in the Irminger Basin in an effort to study middepth convection and recirculation in the basin. This deployment was independent from the extensive release of PALACE floats in the western North Atlantic subpolar gyre led by R. Davis. Two out of six floats were advected in the Irminger Current and entered the Labrador Basin at Cape Farewell. These PALACE floats traveled at an approximate depth of 1500 m with a 14-day cycle during which they spent approximately 12 days at the cruising depth and 1 day at the surface to transmit the data through the satellite ARGOS system. The floats also sample a vertical profile of temperature and salinity during their 4-h ascent. A PALACE float is a quasi-Lagrangian float as it does not follow the exact path of a water parcel. Moreover, its trajectory is contaminated by its ascent, descent, and surface drift.

c. Hydrographic data

Every spring since 1990, the Bedford Institute of Oceanography has been carrying out hydrographic sections between Hamilton Bank, Labrador, and Cape Desolation, Greenland (WOCE line AR7W). In 1996 and 1997, additional hydrographic sections were made for the Labrador Sea Deep Convection Experiment. In October 1996, a hydrographic section was carried out along the AR7W line plus another section in the NW–SE direction close to the Northwest Atlantic mid-ocean ridge direction (Fig. 1). In February–March 1997, the RV Knorr collected an extensive number of hydrographic casts, covering the western part of the Labrador Sea, where the deepest convection was expected to occur. In May 1997, in addition to the AR7W line, three additional sections were done (Fig. 8, center).

d. Current meter data

The mooring B1244 with four Aanderaa RCM-8 current meter recorders was deployed at 55°28.72′N, 53°39.3′W by the C.S.S. Hudson in October 1996. This mooring, located above the 2800-m isobath, was near the core of the deep boundary current. The four instruments were at 201, 1001, 1501, and 2751 m. We also used data collected by five instruments (100, 400, 1000, 2200, and 2485 m) at the mooring M8 deployed in July 1987 at 55°20.82′N, 53°45.3′W.

3. Labrador Sea surface circulation

The surface drifter dataset provides a view of the North Atlantic subpolar gyre’s Lagrangian surface circulation (Fig. 2). A few drifters released in the Newfoundland Basin reached the Northwest Corner of the North Atlantic Current and crossed the Mid-Atlantic Ridge to enter the Iceland Basin. Some drifters released in the Irminger Basin or the Iceland Basin were advected in the East Greenland Current and entered the Labrador Basin. Finally, a few drifters left the Labrador Basin through the shallow Labrador Current to reach the Newfoundland Basin, thus completing the subpolar gyre loop. The drifters released in the European Basin did not leave their region but seemed to converge along the 35°N latitude line, the approximate location of the Azores Current. We notice the large eddy activity compared to the mean flow component of the subpolar gyre. The mean buoy life is 213 days. One buoy had a much longer record of three and a half years, which makes the value of the median 140 days, a more typical buoy life.

Fifty-three drifters drifted through the Labrador Basin (Fig. 3). Thirty-five of the 53 drifters were released in the Labrador Sea itself between 1993 and 1997. Nineteen drifters entered the region at Cape Farewell and were advected in the West Greenland Current. Of these 19 drifters, only 3 went north through Davis Strait. Six
drifters left the current and entered the interior basin. The 10 other drifters followed the bathymetry and turned westward around 61°N. Some of these drifters survived long enough to join the Labrador Current. Five drifters were released north of Hamilton Bank and four others were released south of Hamilton Bank. Six different drifters joined the Labrador Current from the interior Labrador Sea giving a total of 18 drifters that flowed with the Labrador Current.

a. Mean Eulerian circulation

From an Eulerian perspective, the drifter data can be used as an instantaneous picture of the circulation, although it covers an approximate 5-yr period. The time coverage is too sparse to allow any seasonal study. The mean Eulerian picture of the circulation is obtained from the drifters by averaging the data in space and time. Two approaches were considered: bin averaging and objective mapping.

For the first method, all the velocity data available is averaged in bins of 0.5° latitude by 1° longitude. Only the bins with more than 6 buoy-days of data were considered (Fig. 4a). This approach gives an approximate speed of 35 cm s⁻¹ for the shallow West Greenland Current. Just north of the Hamilton Bank, the Labrador Current speed is 20 cm s⁻¹ close to the 1000-m isobath and ~12 cm s⁻¹ above the lower slope. These values are smaller than those obtained by Lazier and Wright (1993) with current meters placed across the Labrador slope. We must note that the core of the classical Labrador Current was undersampled north of Hamilton Bank. More drifters sampled the current south of Hamilton Bank averaging a speed of 30 cm s⁻¹, closer to the 36 cm s⁻¹ obtained by Lazier and Wright. At 700 m, Lavender et al. (2000) observed speeds of the order of 12 cm s⁻¹ for both the West Greenland Current and the Labrador Current. Above the “northern slope,” between the West Greenland Current and the Labrador Current, the surface flow reaches 12 cm s⁻¹ close to the 3000-m isobath and goes up to 20 cm s⁻¹ between the 1000-m and 2000-m isobaths. The flow at 700 m is slower than 5 cm s⁻¹ above the northern slope (Lavender et al. 2000). The interior surface velocity varies spatially and can reach 10 cm s⁻¹.

The second approach consists of objectively mapping the data (Bretherton et al. 1976). First, the velocity data are averaged in bins of 1° latitude by 2° longitude. Then a routine is applied on the mean Eulerian map to obtain a 0.5° latitude by 1° longitude grid by using a Gaussian

![Fig. 4. (a) Map of mean Eulerian velocity deduced by averaging the surface drifter velocity in bins of 0.5° lat by 1° lon. Variance ellipses are drawn at the tip of each arrow. (b) Map of mean Eulerian velocity obtained by objective mapping technique. See text for details.](http://journals.ametsoc.org/jpo/article-pdf/32/2/627/4457223/1520-0485(2002)032_0627_lsbcat_2_0_co_2.pdf)
covariance function with a decorrelation length scale of 150 km (Fig. 4b). The Gaussian acts as a low-pass filter on the data. Both methods yield the same velocity field for the boundary currents. The main difference is found in the interior, where velocities are much weaker (2–3 cm s$^{-1}$) with the objective mapping technique. Offshore of the main boundary currents, a series of recirculation cells constitutes an opposite flow to the boundary currents. The speed in the recirculations is around 3 cm s$^{-1}$. Similar recirculation cells were observed at 700 m using PALACE float data (Lavender et al. 2000). The bin-averaging velocity map displays the same recirculation cells but less clearly defined.

b. Recirculation cells

The Eulerian-averaged data show countercurrents offshore of the principle boundary currents. It is necessary to look at the original drifter trajectories to get more insight into these possible recirculation cells. On the Greenland side (Figs. 5a–e), five drifters were traveling northwestward with the West Greenland Current when they changed direction, left the boundary current and started moving southeastward. The drifters switched direction at different periods of time and at different locations between 59° and 61°N. After a few days drifting in the opposite direction to the West Greenland Current, most of the drifters began meandering in the interior. None of the drifter trajectories showed a complete recirculation loop. At the Eirik Ridge and slightly west of it (Figs. 5a–c and 5f), four drifters stagnated during several months, revealing another possible recirculation region.

On the Labrador side, north of Hamilton Bank (Figs. 6a,b), two drifters joined the Labrador Current after having drifted in the opposite direction. South of Hamilton Bank (Figs. 6c,d), four drifters left the Labrador Current at different times in a similar fashion to what was observed above the Greenland slope. The trajectory in Fig. 6d is the clearest evidence of a current parallel and opposite to the Labrador Current. Figures 6d and 6f suggest that the northward flow could be part of a large North Atlantic Current loop as the drifters eventually head eastward toward the Charlie-Gibbs Fracture Zone. However, there has never been any report of the North Atlantic Current Northwest Corner reaching as far north as 55°N (Lazier 1994; Kearns and Rossby 1998). The North Atlantic Current is usually considered to be confined within the 4000-m isobath. Another interpretation would be that after flowing with the current opposite to
the Labrador Current, the drifters were advected in a current exiting the Labrador Basin that ends up joining the North Atlantic Current farther east. The offshore part of the recirculation cells is far enough out to be in the area of weak stratification of the Labrador Sea, explaining why the same features are clearly observed both at the surface with surface drifters and at 700 m with PALACE floats (Lavender et al. 2000).

c. Kinetic energy

A map of total kinetic energy of the Labrador Sea deduced from the surface drifters (not shown here) emphasizes mostly the shallow West Greenland Current and the shallow Labrador Current with values up to 700 cm$^2$ s$^{-2}$. Figure 7 shows a map of eddy kinetic energy $[0.5(u'^2 + v'^2)]$ deduced from the surface drifter data where the fluctuation velocity is calculated by subtracting the Eulerian mean within each bin. The Northwest Corner of the North Atlantic Current shows values from 700 to 200 cm$^2$ s$^{-2}$ at the outer limit. In the Labrador Sea, two maxima of eddy kinetic energy with values up to 400 cm$^2$ s$^{-2}$ are found close to the Greenland slope. The first one, located above the Greenland slope at 60.5°N, 50°W, in the middle of the West Greenland Current, may be the result of the interactions between the current and the interior described in the preceding section. The second one is located slightly west of the point where the 3000-m isobath curves and separates from the Greenland slope. Sea surface brightness satellite images reveal that numerous eddies are formed in this area (Prater 2002). The curvature of the bathymetry and the change in slope may affect the stability of the current and be responsible for this zone of high eddy kinetic energy. A pool of high eddy kinetic energy is found at 700 m at the same location (K. Lavender 1999, personal communication). White and Heywood (1995) computed maps of eddy kinetic energy for the North Atlantic sub-

### Table 1. Transit time and speed for the four segments defined in Fig. 3.

<table>
<thead>
<tr>
<th>Segment</th>
<th>Transit time (days)</th>
<th>Mean speed (cm s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mean ± std (median)</td>
<td></td>
</tr>
<tr>
<td>Segment 1</td>
<td>21.3 ± 7.49 (19.2)</td>
<td>32.2</td>
</tr>
<tr>
<td>Segment 2</td>
<td>49.3 ± 8.9 (45.5)</td>
<td>18.9</td>
</tr>
<tr>
<td>North branch</td>
<td>114 ± 40.3 (40.3)</td>
<td>17.9</td>
</tr>
<tr>
<td>South branch</td>
<td>87.4 ± 29.4 (85)</td>
<td>10.6</td>
</tr>
<tr>
<td>Segment 3</td>
<td>37.6 ± 15.6 (37.3)</td>
<td>18.5</td>
</tr>
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polar gyre using the altimetric data from TOPEX/Poseidon and Geosat. They also found a maximum of eddy kinetic energy in the Labrador Sea at the same location at all seasons.

d. Transit times

Mean speed and travel time were estimated for four segments of the boundary currents (Table 1). The four segments are delimited by the five black lines in Fig. 3 (Segments 1–4, going counterclockwise from the Greenland coast). Only those drifters that traveled through an entire segment were considered in estimates of transit times and mean speeds. The Segment 1 mean speed of 32.2 cm s\(^{-1}\) is in good agreement with the speed measured by Clarke (1984) at Cape Farewell by an instrument 100 m deep above the 1000-m isobath. The drifters show a reduction in speed by about a factor of 2 when they reach the northern slope of the basin. The drifter tracks suggest two main flow paths above the northern slope between Greenland and Labrador. Mean speeds for the two paths are close, but the drifter tracks show that the southern branch meanders more than the northern branch, thus giving different transit times. These two paths are described in more detail in the following section. The two segments covering the southeastward flowing Labrador Current give an average speed of 14.5 cm s\(^{-1}\). Most drifters traveled in waters deeper than 2000 m and missed the core of the shelfbreak Labrador Current close to the 1000-m isobath. This value is more representative of the “deep” Labrador Current (Lazier and Wright 1993).

Transit times are important for considering the impact of winter convection on the boundary current. The drifter transit times suggest that a water mass would take approximately 185 days (using the median values) to go from Cape Farewell to the 50\(^\circ\)N latitude line on the Labrador side. Between the two extremities of the AR7W line, the transit time is 147 days.

4. Horizontal and vertical structure of the currents

a. West Greenland and Labrador Currents

Using the drifters advected in the boundary currents at different times and the four hydrographic sections made in May 1997, the horizontal and vertical structure of the boundary currents were examined. At Cape Farewell (Fig. 8a), the drifters show the shallow West Greenland Current with a maximum speed of 90 cm s\(^{-1}\) on the shelf. The current speed decreases toward the 3000-m isobath with values close to 20 cm s\(^{-1}\) at the 2000-m isobath. At Cape Desolation (Fig. 8b), the current maximum speed is around the 2000-m isobath and reaches 95 cm s\(^{-1}\). The current is slower on both sides of the 2000-m isobath with speeds around 35 cm s\(^{-1}\) above the shelf and 25 cm s\(^{-1}\) close to the 3000-m isobath. At the northern section on the Greenland side (Fig. 8c), the shallow West Greenland Current slows to 70 cm s\(^{-1}\). At the northern section above the Labrador slope (Fig. 8e), the speeds range between 25 and 35 cm s\(^{-1}\). North of Hamilton Bank (Fig. 8f), where the Labrador slope is steeper, one drifter had a speed of 65 cm s\(^{-1}\) between the 1000-m and 2000-m isobaths. This value is close to the instantaneous speed estimated by Lazier and Wright (1993) for the shelfbreak Labrador Current. However, the drifters mostly sampled the lower slope part of the boundary current where speeds range between 25 and 35 cm s\(^{-1}\). At the southern section on the Labrador side (Fig. 8g), the boundary current slows down with two maxima at the 1000-m and 2000-m isobaths. Between these two isobaths, the drifters display speeds around 12 cm s\(^{-1}\).

Using the hydrographic data, the speed difference between 18 m, the depth of the shallowest record, and 1000 m was computed. The resulting vertical shear is compared with the drifters assuming that the drifter instantaneous velocity is representative of the flow when the sections were occupied. On the Labrador side, there is a maximum vertical shear at the 1000-m isobath and a minimum close to the 2700-m isobath. The shear is almost zero in every panel at the 2700-m isobath. However, the maximum shear varies in amplitude between sections. At the southern and northern section (Figs.
The shear reaches 10 cm s\(^{-1}\) at the 1000-m isobath. It goes up to 50 cm s\(^{-1}\) at the AR7W section (Fig. 8f) such that the difference in shear between the shelf break and the lower slope reaches an order of magnitude. This structure matches well with the description of the Labrador Current given by Lazier and Wright (1993) with a baroclinic shelfbreak current and a more barotropic lower slope current, the deep Labrador Current. Lazier and Wright observed the relatively barotropic current with the mooring M8 deployed north of Hamilton Bank at the 2500-m isobath in July 1987 and recovered in July 1988 (Fig. 1). The average speeds measured by current meters at M8 are 16, 19, 19, 16, and 14 cm s\(^{-1}\), respectively, at 100, 400, 1000, 2200, and 2485 m. Here, from hydrographic data, the minimum shear is found at the 2700-m isobath, but the low number of stations above the Labrador slope does not allow one to define precisely the center of the deep Labrador Current. Mooring B1244 located at the 2800-m isobath also sampled a barotropic flow. The 20-month average speeds recorded by B1244 are 20.6, 17.2, 16.3, and 24.4 cm s\(^{-1}\), respectively, at 200, 1001, 1501, and 2751 m. The speed
Fig. 9. Potential temperature (°C) section and potential density \(\sigma_0\) contours from the May 1997 NW–SE hydrographic section (contour interval is 0.05 kg m\(^{-3}\)). The circled stars indicate where surface drifters drogued at 15 m crossed the northern end of this hydrographic section (the stars are plotted at 50-m depth for clarity). The crossing locations of the 2000-m and 3000-m isobaths are indicated. The distance (km) is from the northernmost station at 63.42°N, 58°W (see Fig. 1).

Fig. 10. Potential temperature (°C) in the 27.55–27.6 \(\sigma_0\) layer objectively mapped from the Aug–Oct 1965 hydrographic cruise data. Stars indicate the location of the stations where data points were found in the layer of interest. Circles indicate the location of all the stations for the cruise.

is almost uniform over the top 1500 m but the deep western boundary current sampled at the bottom increases the vertical shear in the column. The speed averaged over the top 1500 m is \(18 \pm 2\) cm s\(^{-1}\). The similarities between the records of the moorings M8 and B1244 suggest that they were located symmetrically on both sides of the core of the deep Labrador Current.

On the Greenland side (Figs. 8b,c), the maximum vertical shear is found between the 2000-m and 3000-m isobath and can reach 45 cm s\(^{-1}\). Generally, the shear on the Greenland slope exceeds the shear on either the northern slope or the Labrador slope. Similarly to the Labrador Current, the maximum shear in the shallow West Greenland Current matches with the location of the maximum speed and the shear is weaker close to the 3000-m isobath. However, the Greenland slope being steeper than the Labrador slope, the relatively baroclinic West Greenland Current extends above the lower part of the continental slope and may mask partially the presence of a more barotropic current.
b. Bifurcation of the West Greenland circulation

It appears that when the cyclonic boundary current system encounters shoaling depths on the Greenland side, it produces a bifurcation where a fraction of the flow follows the Greenland coast and the rest heads southwestward. Figure 8d gives the horizontal and vertical structures of the currents above the northern slope of the Labrador Basin. The drifters are distributed into two groups. One group of drifters is confined between the 1000-m and 2000-m isobaths and the second group is dispersed around the 3000-m isobath. This distribution is also visible in the complete set of drifters (Fig. 3), where there is a significant absence of trajectories...
FIG. 12. Trajectories of PALACE floats 77 (a) and 78 (b). Black lines indicate the deep displacement of the floats. Gray lines indicate the surface displacement of the floats. Stars indicate the locations of the vertical profiles sampled by the floats. A few profile numbers are written.

FIG. 13. $\theta$-$S$ diagram showing PALACE 78 profiles 41 and 44 and a profile from the interior Labrador Sea collected in Jun 1998, averaged over a few isopycnal layers surrounding the ISW. The result of along-isopycnal mixing between profile 41 and the interior profile is plotted with a dashed line marked with squares.
between the 2000-m and 3000-m isobaths. The northern group of drifters displays speeds between 5 and 30 cm s\(^{-1}\), with the maximum found at the 2000-m isobath. The southern group has speeds around 25 cm s\(^{-1}\). The current speed above the northern slope is one-fourth of the shallow West Greenland Current speed at Cape Desolation. The drifters included in the northern group had well-defined trajectories, whereas the southern group followed more meandering paths. The NW–SE potential temperature section done in May 1997 (Fig. 9) shows that each group of drifters is associated with a similar type of water column made of warm ISW water between 200 and 500 m and cold Greenland shelfbreak water at the surface. The vertical shear in the top 1000 m above the northern slope (Fig. 8d) is an order of magnitude smaller than above the Greenland slope. It is minimum around the 2500–2800 m isobaths, giving a vertical shear distribution similar to the distribution above the Labrador slope.

It is possible to check the spatial distribution of warm water using the only extensive survey of the Davis Strait area, done from August to October 1965. We objectively mapped potential temperature in the 27.55–27.6 \(\sigma_0\) layer from the 1965 hydrographic stations (Fig. 10). There is a distinct warm ISW flow between the 2000-m and 1000-m isobaths that joins the Greenland and Labrador slope. The warm water mass sampled by the southernmost section close to the 3000-m isobath suggests that there are two main paths for the ISW above the northern slope.

Reynaud et al. (1995) obtained two currents above the northern slope in their summer-mean circulation map deduced by inverse methods applied to objectively mapped temperature and salinity fields. The temperature and salinity fields were obtained from the hydrographic
data available since 1910. A distinct current above the northern slope along the 3000-m isobath was also found in the mean circulation obtained by Tang et al. (1996) with a linear three-dimensional model. Considering the location and the vertical structure of these two currents, it seems reasonable to infer that the northern branch is an extension of the shelfbreak component of the West Greenland Current and the branch closer to the 3000-m isobath is part of the barotropic deep Labrador Current. Hereafter, we will refer to the northern current above the northern slope as the West Greenland Current Extension (WGCE).

5. Irminger Sea Water evolution in the Labrador Basin

The Irminger Sea Water inflow in the Labrador Sea significantly influences the final product of convection in the western Labrador Sea. We analyzed the data collected by two PALACE floats that approximately followed the path of the ISW in the Labrador Sea. The temperature and salinity regularly sampled by these floats allowed us to track the evolution of the ISW around the basin. We also analyzed the significant seasonal and interannual variability in ISW properties displayed in hydrographic and mooring data.

a. Mixing of the Irminger Sea Water

Hydrographic sections along the AR7W line show that the amount of warm and saline water is smaller on
the Labrador side than on the Greenland side (Fig. 11b). The ISW maximum temperature and salinity in May 1997 differed between both slopes by 0.9°C and 0.045 psu, suggesting that the ISW is cooled and freshened as it travels around the basin.

The two PALACE floats (77 and 78) traveled at 1500 m, such that their deep velocities are weaker than the velocities at the ISW level around 300 m. The two PALACE floats took 168 and 294 days to travel between both ends of the AR7W line following, respectively, the West Greenland Current Extension and the deep Labrador Current above the northern slope. The transit time of the surface drifters along the same path (147 days) gives the upper limit for the possible travel time of the ISW layer (200–500 m) because of the flow being surface intensified. The floats that traveled through the Labrador Sea in summer–fall of 1998 and winter–spring of 1997/98 suggest different regimes for the two periods.

PALACE float 78 entered the Labrador Basin on 10 May 1998 and was just north of Hamilton Bank on 8 November 1998 (Fig. 12b). The θ–S diagrams in Figs. 13 and 15 show the transformation of ISW during the summer-fall period using PALACE float 78 profiles 41,
FIG. 18. Potential temperature (a) and salinity (b) averaged over a few isopycnal layers surrounding the ISW from PALACE float 77 profiles 24–51 collected in the Labrador Basin. (c) Potential density $\sigma_0 (\text{kg m}^{-3})$ section from PALACE float 77 profiles 24–51. There is a brief period with well-mixed properties as the float approached the Labrador slope (59°–61°N, 58°W).

The ISW temperature and salinity maximum goes from 4.75°C, 34.94 psu at the Greenland slope to 4.675°C, 34.91 psu above the northern slope, and 4.28°C, 34.85 psu at the Labrador slope. Comparison of the section in Fig. 11b and the locations of profiles 41 and 53 above the continental slopes suggests that these temperatures are representative of the ISW. It is more difficult to judge the representativeness of profile 44 as the surface drifters showed that the deep Labrador Current meanders above the northern slope.

Figure 13 suggests that along-isopycral mixing with subsurface interior water is a good candidate for the ISW transformation. We considered the result of along-isopycnal mixing between profile 41 of PALACE float 78 (water column A) and an interior CTD profile collected in July 1998 (water column B) using the following formulas:

$$Z_v = (1 - b)Z_a + bZ_b$$

$$\theta_v = (1 - b)\theta_a \left(\frac{Z_a}{Z_c}\right) + b\theta_b \left(\frac{Z_b}{Z_c}\right)$$

$$S_v = (1 - b)S_a \left(\frac{Z_a}{Z_c}\right) + bS_b \left(\frac{Z_b}{Z_c}\right)$$

The variables are considered for specific $\sigma_0$ layers, with $Z(\sigma)$ the thickness between the isopycnals $\sigma$ and $\sigma + \Delta\sigma$, and $b$ the mixing coefficient. The ISW temperature and salinity maximum above the northern slope is obtained with $b = 0.22$, giving the fraction of surface interior water involved in the mixing product. Using
this same value for $b$ does not allow one to reproduce as well the water surrounding the ISW temperature and salinity maximum. This suggests a variable mixing ratio throughout the water column or the effect of other processes. For instance, the shallower water may also be sensitive to surface heat fluxes.

If we assume that along-isopycnal mixing with surface interior waters is the main cooling process of the ISW, Fig. 9 shows that above the northern slope the southern branch would be cooled before the northern branch. Starting with profile 46, float 78 left the deep Labrador Current and joined the shelfbreak Labrador Current. Coincidently, the float sampled colder and fresher waters in the isopycnal range of the ISW (Fig. 14). This suggests that the ISW carried by the WGCE can mix along isopycnals with the fresh and cold water of the southward flowing Baffin Island Current. We apply the same mixing model as described above but with water column B being a profile taken in the Baffin Island Current during the Hudson 1965 cruise. It seems unrealistic to use such an old CTD profile, but we suppose it is good enough to support the conceptual idea of mixing process that we want to present here.

Figure 10 shows that the ISW carried by the WGCE can mix along isopycnals with the fresh and cold water of the ISW carried by the WGCE. Figure 15 shows that the ISW temperature and salinity maximum in profile 53 can be obtained by along-isopycnal mixing with $b = 0.01$. The small value for $b$ is due to the large temperature and salinity difference between the two water masses. However, the fresh and cold Baffin Island Water has only isopycnals in common with the ISW down to approximately $27.6 \sigma_0$. This range of isopycnals covers only the upper part of the ISW core and it is unclear how the lower part of the ISW core carried by the WGCE could be cooled and freshened significantly.
Applying a simple one-dimensional diffusion model ($\theta_s = \kappa \theta_{ss}$) to profile 41 shows that diapycnal diffusion plays a minor role in the cooling and freshening of the ISW. With $\kappa = 10^{-4} \text{ m}^2 \text{ s}^{-1}$, diffusion reduces the warmest temperature of the ISW by only $0.03^\circ \text{C}$ and the salinity by 0.01 psu after 147 days (Fig. 15). So we conclude that, the ISW being cooled differently in the northern and southern branch during the transition between Greenland and Labrador, we may observe two distinct ISW products above the Labrador slope (Fig. 16).

PALACE float 77 entered the Labrador Basin on 28 September 1997 and reached Hamilton Bank on 27 September 1998 (profile 48 in Fig. 12a). Figure 17 displays the $\theta$-$S$ diagrams for a few profiles taken by the PALACE float 77 in the Labrador Sea region during the winter–spring period. The first profile was collected along the Greenland slope in December 1997 and displays a warm and salty peak at 600 m. The following profiles show the warm peak being cooled and freshened by mixing. The profiles 33, 34, and 36 (profile 35 was not recovered) indicate intrusions of cold and fresh water. The intrusions are concentrated along the isopycnals 27.6 $\sigma_t$, 27.55 $\sigma_t$, and 27.68 $\sigma_t$. The recent winter NW–SE hydrographic section done in February 1997 (Fig. 11a) shows that the isopycnals corresponding to the ISW above the northern slope outcrop southward in the center of the basin and thus are directly forced by the atmosphere. The cold water intrusions are accompanied by a slow erosion of the surface stratification.

Figure 18 looks at the evolution of the water column from an isopycnal point of view. Some points are missing along the curves for several reasons: First, profiles 31 and 35 were not recovered. Second, profile 28 was only recovered down to 500 m. Finally, no point was drawn when one of the limits of an isopycnal layer was not present in the water column. The top two layers represent approximately the surface layer, and the four others correspond to the ISW layer. Up to profile 34, most of the layers are progressively cooled and freshened with the surface layers being the most affected. At profile 36, the two top layers are no longer observed, and the lower layer temperatures are homogenizing. Profile 37 collected on 26 April 1998 sampled a water column uniform over 400 m ($3.5^\circ \text{C}, 34.8$ psu; Fig. 18). Beginning with profile 39, we observe the warming of the surface and the progressive restratification of the water column as the float traveled closer to the 2000-m isobath.

Assuming that profile 37 is the product of convection taking place along the boundary current, we attempted to reproduce it by removing buoyancy from profile 30 collected above the Greenland slope. Using the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis heat fluxes corrected with respect to in situ data (Renfrew et al. 2002) for the corresponding period, the result is a uniform water column 153 m deep. This suggests that the cooling of the ISW layer by along-isopycnal mixing is a necessary preconditioning process to convection along the boundary current.

b. Interannual and annual variations

Figure 19 shows the variations in temperature and salinity of the Irminger Sea Water on both ends of the AR7W line through the 1990s. On the Greenland side, the maximum temperature and salinity within the density range 27.6–27.75 $\sigma_t$ were used to characterize the ISW. On the Labrador side, we averaged the temperature and salinity between 27.6 $\sigma_t$ and 27.75 $\sigma_t$, for each profile above the slope. Then we chose the maximum average temperature and salinity as the characteristic temperature and salinity for the ISW. The reason for this approach will become clear when we look at the seasonal variations. The ISW potential temperature and salinity on the Greenland side does not present any clear trend through the decade. The LSW cooled and freshened until 1994, revealing its renewal by deep convection, then became warmer and saltier since then. The water in the ISW density range on the Labrador side varied similarly to the LSW except that it kept cooling and freshening until 1995. There is no certainty about these trends because some points are missing (no data in May 1992 and July 1998 on the Greenland side) and some other data were obtained from stations close to...
Fig. 21. $\theta$-$S$ diagram showing CTD profiles collected above the Greenland slope (black) and the Labrador slope (gray) in (a) Oct 1996 and (b) May 1997.

the 3000-m isobath, missing the core of the ISW (May 1993 and July 1999). However, the winters 1990–95 were much colder than those of 1996–99, and this may reflect the warming trend.

The variability in the ISW temperature and salinity difference between the Greenland and Labrador slopes is large. [For the period 1990–99, we found mean ($\Delta T$) $\pm$ std($\Delta T$) = 1.1 $\pm$ 0.4 and mean ($\Delta S$) $\pm$ std($\Delta S$) = 0.08 $\pm$ 0.04].

The sections from October 1996 and February 1997 added to the May 1996 and May 1997 sections give a rough picture of the seasonal variation (Fig. 20). The potential temperature time series at 200 m from the mooring B1244 (see Fig. 1), located at the 2800-m isobath on the Labrador slope from October 1996 to June 1998, agrees with the seasonal variations from the CTD. The mooring time series shows that the potential temperature on the Labrador side kept rising to 4.5°C until December and then decreased to 3.2°C in March. The ISW seems to vanish from the Labrador slope in winter when the potential temperature recorded at 200 m reaches the values recorded at 1000 m. The speed of the current along the Labrador slope (see Section 3 and 4) suggests that this is the result of convective activity upstream of the mooring location more than a local process (Pickart et al. 1997). Accordingly, Fig. 21 shows that in May 1997, barely any warm and salty water was observed in the range 27.55–27.75 $\sigma_t$ above the Labrador slope, explaining why the ISW layer on the Labrador side is shown fresher than the LSW in Fig. 19. This was also observed in most of the late spring AR7W sections in the second half of the 1990s. The temperature recorded by the mooring increased gradually through summer and fall to reach again values more characteristic of the ISW. This gradual recovery emphasizes the wide extent of the convective activity and its accompanying isopycnal mixing. The transit times given in Section 3 show that during the seven month warming period water 200–500 m deep traveled around almost the entire Labrador Sea.

6. Summary

Using surface drifters released in the North Atlantic subpolar gyre and several hydrographic sections from the Labrador Sea Deep Convection Experiment, we have described the boundary circulation of the Labrador Basin. The drifters show a pronounced asymmetry between the boundary currents on the Greenland and the Labrador side: The flow is rapid and concentrated above the steep Greenland slope, whereas it is broader and slower above the Labrador slope. The surface drifters, as well as the deep floats, indicate that the shallow West Greenland Current significantly slows around 62°N
where the slope decreases as the 3000-m and 2000-m isobaths curve westward. In the transition region between the Greenland and Labrador slopes, the currents have, on average, half the speed of the shallow West Greenland Current. Then, the flow joins the fast Labrador Current as the 3000-m comes closer to the 2000-m isobath. This close relation between speed and bottom topography reflects the fact that water columns tend to follow $f/H$ contours, $f$ being the Coriolis factor and $H$ the bottom depth.

Offshore of the cyclonic boundary circulation, counterflows with speeds around 3 cm s$^{-1}$ appear in Eulerian averages as well as individual trajectories. In agreement with the work of Lazier and Wright (1993), we found that the circulation above the Labrador and the northern slope is made of two main components: a more baroclinic current whose center is located between the 2000-m and 1000-m isobaths, the Labrador Current and the West Greenland Current Extension, and a more barotropic component centered between the 2000-m and 3000-m isobaths, the deep Labrador Current. There is evidence of a similar current structure above the Greenland slope but it is not as clear as for the northern slope or the Labrador slope.

We have analyzed the data from two PALACE floats that tracked the ISW during the summer–fall of 1998 and the winter–spring of 1997/98. In this brief sample, the ISW carried by the deep Labrador Current was cooled and freshened by along-isopycnal mixing with subsurface interior water. The recirculation cells observed both here and by Lavender et al. (2000) may be involved in the mixing process. We suggest that the ISW carried by the WGCE above the northern slope is cooled and freshened by along-isopycnal mixing with Baffin Island Water.

PALACE float 77, which traveled in the shallow West Greenland Current during winter, did not sample a second site of deep convection close to the Greenland coast as suggested by Tang et al. (1999). We believe that the second site of uniform water that Tang et al. (1999) observe in the data (their Fig. 20) is the branch of the “deep” Labrador Current that leaves the Greenland coast carrying away some Irminger Sea Water. Nevertheless, the same float sampled a uniform water column 400 m deep in the northwest part of the basin where the highest air–sea heat fluxes occur during winter (Fig. 22). A hypothesis for the fate of the Irminger Sea Water in winter is suggested. The ISW mixes along isopycnals with interior surface waters directly forced by the atmosphere (Fig. 11a). This happens predominantly above the northern slope of the basin, where the currents are slower. During the 49–114 days (Table 1) needed by the surface waters to cross between Greenland and Labrador, the large heat fluxes may erode the fresh and cold water cap covering the ISW. Though the PALACE float record suggests Lagrangian convection, we cannot disregard the possibility that the fresh cap is eroded while the waters are trapped in a recirculation cell in the northwest part of the basin. For example, PALACE float 77 stayed almost three months between 58° and 60°N (profiles 37–43). The uniform water column sampled by the PALACE float was inshore of the 3000-m isobath and may have been advected by the slope current. This could play a role in the formation of the 1000-m uniform water column observed in winter above the Labrador slope between the 2000-m and 3000-m isobaths in the path of the deep Labrador Current (Pickart et al. 2002). More Lagrangian sampling of the convection taking place above the northern slope will be necessary to fully understand the specificity of convection above the Labrador slope.

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