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Labrador Sea Water Circulation in the Northern North Atlantic Ocean

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THE FLORIDA STATE UNIVERSITY
COLLEGE OF ARTS AND SCIENCES

LABRADOR SEA WATER CIRCULATION IN THE NORTHERN NORTH
ATLANTIC OCEAN

By

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ABSTRACT

Direct observation of velocities with drifting floats near 1500m depth in the North Atlantic Ocean north of about 45°N are combined to derive a new basin-wide mean circulation within the layer of Labrador Sea Water. Inverse methods are used to determine a basin-scale circulation constrained by non-divergence and conservation of climatological large-scale potential vorticity, with the aim of investigating large-scale potential vorticity dynamics, improving circulation estimates near the basin boundaries, and determining the relative amount of recirculation in sub-basin gyres with (aliased) direct velocity data. Eulerian averages are created from combined datasets and a streamfunction of the non-divergent flow is computed from the velocity curl and the imposed boundary conditions. Unlike deep flow regimes at low latitudes, where the mid-depth interior circulation is dominated by zonal currents, the mid-depth subpolar flow is dominated by gyres.

As suggested by previous estimates of mid-depth circulation, the Labrador Sea Water layer is rapidly circulating around the periphery of the subpolar basin in northern and western boundary currents. Recirculation within basins and exchange between the Labrador Basin and Irminger Basin is modified by the large-scale PV constraints. Eastward flow across the Mid-Atlantic Ridge is concentrated in the Charlie Gibbs Fracture Zone with return flow above the Reykjanes Ridge. A weaker subtropical return flow around the Azores Plateau appears, but data are insufficient to map adequately this region.

CHAPTER 1

INTRODUCTION

Labrador Sea Water (LSW) is a water mass formed in the Labrador Sea that occupies a large part of the mid-depth North Atlantic Ocean. It is characterized by its fresh, cold, low stratification or low values of large-scale potential vorticity (PV) and high concentration of anthropogenic tracers. LSW is renewed during wintertime deep convection, a process sensitive to oceanic and atmospheric conditions (Lab Sea Group, 1998). The essential atmospheric condition is strong buoyancy forcing, associated with the dry and cold winds blowing in the area. Oceanic conditions associated with deep convection are a weakening of the stratification near the surface and closed circulation such that near-surface water is trapped for a longer period of time while undergoing surface forcing. LSW is the final product of the successive mode waters defined by Talley and McCartney (1982) and Talley (1999). From the formation area, the LSW spreads throughout the North Atlantic and spatial maps of its signature show three main spreading tongues (Fig. 1.1). One route is northeastward into the Irminger Sea. Another branch heads east within the North Atlantic Current and crosses the Mid-Atlantic Ridge (MAR). The third one is within the Deep Western Boundary Current. These pathways were highlighted for example by maps of lateral distribution of the mid-depth LSW potential vorticity minimum (Talley and McCartney, 1982), or maps of CFC distribution (Rhein et al., 2002). East of the MAR, Cunningham and Haine (1995) showed that LSW flows northward into the Icelandic Basin and the Rockall Trough as well as southward above the MAR.

Boundary currents and sub-basin scale recirculation were observed by Lavender et al. (2000), who constructed an average circulation using Profiling Autonomous Lagrangian Circulation Explorer (PALACE) float data. Their floats were released in the western subpolar North Atlantic at nominal depths of 400, 700 and 1500 m and velocities were corrected to 700 m using climatological thermal wind (Levitus) [Note that 700m is not the

depth of Labrador Sea Water outside of the convection zone, but that the vertical shear is weak and corrections were generally small over most of the region]. The recirculation structures were organized into cyclonic gyres in the Irminger Sea and the Labrador Sea. Lavender et al. (2003) mapped circulation at 1500m depth from a subset of floats drifting near 1500m depth, but many of these floats remained in the Labrador Sea. Lilly et al. (2003) questioned the representativeness of the averaged float data in the presence of small-scale eddies thought to be active during the period of the experiment; this, together with the vertical extrapolation to 700m helps to motivate a reanalysis of all the observations near 1500m.

The effect of recirculation on tracers in the subpolar gyre was investigated by Straneo et al. (2003) who simulated the spreading of LSW with an advective-diffusive model, whose mean flow and eddy diffusivity were derived from the same data used in Lavender et al. (2000). They concluded that advection and diffusion were comparable, and that advection governed export out of the Labrador Sea. The strength of recirculation governs the residence time of tracers in the region.

Independently, Bower et al. (2002) mapped the LSW circulation mainly on the east side of the MAR using isopycnal and isobaric floats data near 1500m depth (their Fig. 3b). They depicted a cyclonic recirculation in the Iceland Basin with a strong flow through the Charlie Gibbs Fracture Zone (CGFZ) corresponding to the deep extension of the NAC. South of 52°-55°N, they show a broad but weak southward flow, poorly resolved away from topography because of relatively strong eddy activity. Their analysis of the flow suggested that there was a current across the MAR both in the NAC and to south of it. However, their study also suggests that the strong flow near the CGFZ originates mostly from the northern cyclonic recirculation, implying a weak net exchange of LSW between the western and eastern sides of the ridge beneath the NAC. This result might depend to some extent on their mapping procedure, which did not conserve mass along the north-east boundary of the Iceland Basin. In the present study we combine data from the various experiments in the North Atlantic in order to construct a complete and consistent analysis of the float data near 1500m depth, to investigate the mean circulation and recirculation within basins. A no-flow boundary constraint is imposed strongly to maintain boundary currents, and climatological

Figure 1.1: Potential vorticity of the Labrador Sea Water, defined as f/h where h is the thickness between the two neutral density layers defining LSW. The contour interval is $0.2 \times 10^{-7} \text{ 1/(ms)}$. The isobaths are 1500m and 2500m.

potential vorticity (PV) constraints are used to attempt to overcome some of the bias from small-scale eddies.

CHAPTER 2

METHOD AND DATA

2.1 Approach

Our objective is to combine several datasets (Lavender et al. (2000), Lavender et al. (2003), Bower et al. (2002)) in order to map the LSW circulation at 1500-1750 m depth. To obtain a representation of the LSW circulation, we proceed in three steps. First, spatiotemporal average velocities and associated errors are computed. The mean fields are then combined on a regular grid by averaging velocities and errors in the overlapping bins. Second, inverse methods are applied to constrain the flow by assuming geostrophy on a f -plane and a homogeneous LSW layer contained between two neutral density surfaces. Third, a streamfunction is calculated from the new velocity estimates.

To aid mapping and to bring dynamics to bear we use a method for computing the streamfunction which attempts to be consistent with the potential vorticity distribution of the Labrador Sea Water. Both barotropic (f/H , where f is the Coriolis parameter and H is the full ocean depth) and local PV (f/h , where h is a layer thickness) are considered. In principle, the degree of misfit to these constraints is a measure of PV sources or dissipation; not unexpectedly, these residuals proved to be too noisy to interpret with the present analysis.

2.2 Data

The datasets are listed in Table 2.1. Fig. 2.1 shows the displacements and trajectories of the floats used for the geographic bin average of the velocities. These float trajectories span the northern North Atlantic Ocean, north of about 45°N at a nominal depth of 1500m. The float data were all collected from the period 1996 through 2001. North of 45°N there are a few areas with negligible sampling: the western boundary current region in the southwest, part

Table 2.1: Summary of the datasets used in the circulation estimates. Acronyms are the US studies Atlantic Circulation and Climate Experiment (ACCE), Labrador Sea Deep Convection Experiment (LSDCE), and the European Studies of the northern North Atlantic (SFB and EUROFLOAT).

experiment	float type	depth	period	nb. floats	floatyears
ACCE	PALACE	1500	October 1996 - December 2000	42	125
LSDCE	PALACE	1500	March 1997 - October 2001	15	27
EUROFLOAT	MARVOR	1750	September 1996 - June 2003	21	72
SFB	RAFOS	1500	May 1997 - August 2000	34	47

of the Irminger Sea, and the Bay of Biscay. [Note that the coverage is different at shallower depths, see e.g. Bower et al. (2002)]. From the representation of the float trajectories on Fig. 2.1, a few major features of the ocean circulation are apparent, including northern and western boundary currents, and flow around the Reykjanes Ridge. In the eastern trough, there is a southward and westward drift of the floats south of 45°N passing around the MAR and the Azores Plateau. In general, there is a contrast between the flow of the narrow boundary currents and the turbulent character of the interior circulation at these depths.

2.3 Mean velocities

Considering the flow as being approximately homogeneous and stationary, the domain is divided into bins of 1°latitude by 2°longitude. Within each bin, the mean velocity components $\langle u_i \rangle$ are calculated, as well as u'_i , the departure from $\langle u_i \rangle$ (Fig. 2.2). The floats used in our data set fall into two categories: floats measuring subsurface displacements at periods ranging from 4 to 20 days (PALACE) and acoustically tracked floats measuring trajectories at one day interval (MARVOR and RAFOS). Because these two float types give different measures of the variance, we computed average velocities and standard errors for each of them separately. The averaging of the PALACE velocities was weighted by the subsurface periods (Davis, 1998). At the overlapping bins (i.e. bins containing both float types), the two mean fields and squared errors were averaged.

Following Krauss and Böning (1987) the associated standard error is defined as

$$\epsilon = \frac{2\sqrt{\langle u_i'^2 \rangle}}{\sqrt{N_{dof}}} \quad (2.1)$$

Figure 2.1: Trajectories of the subsurface floats set at 1500 dbars (PALACE and RAFOS types) and 1750 dbars (MARVOR type). The isobaths are 500m, 1500m, 2500m and 3500m.

where $\langle u_i^{l^2} \rangle$ is the standard deviation and N_{dof} the degree of freedom (statistically independent data in the box considered). The standard error gives the 95% confidence interval according to the Student-t test (e.g. Krauss and Böning, 1987). The degree of freedom N_{dof} is defined as

$$N_{dof} = \frac{N \times \Delta t}{\tau} \quad (2.2)$$

where Δt is the sampling time, N the number of observations in a box for one float, τ is the Lagrangian integral time scale. The integral time scales were calculated from single

Figure 2.2: Bin averaged velocities after combining averaged velocities from PALACE, RAFOS and MARVOR floats. Standard error ellipses are centered on the origins of the velocity vectors.

floats and averaged within a bin. The integral time scale was computed from the first zero crossing of the autocovariance function. For the PALACE floats, a integral time scale of 10 days was assumed.

2.4 Inverse Model

The bin average velocity field computed above gives a measure of the mean circulation at the LSW level. In this section, we want to combine this measure in a simple inverse model

in order to obtain a more realistic estimate of the circulation, i.e. a circulation satisfying physical constraints. The constraints we chose are approximate f-plane geostrophy hence non-divergence of the (2-D) flow and conservation of potential vorticity. The floats used here are neutrally buoyant and do not follow precisely levels of neutral density. However, they do sample well the bulk LSW layer. Fig. 2.3 shows a hydrographic section running through the subpolar North Atlantic ocean. The nominal floats depths of 1500 m and 1750 m lie inside the depth range of LSW, identified on the figure by its potential vorticity minimum. Later on, we will assume that the flow is depth independent between these two neutral density surfaces.

Model formulation. The finite difference form of the constraints is written on the grid defined by the space-time average circulation previously calculated. These vectors are here considered to be an estimate of the Eulerian circulation, and are noted as lower case u and v . We want to obtain a new velocity field upper case U and V (see Fig. 2.4), the unknowns of the problem, defined by

$$\begin{aligned} U_{i,j} + n_{i,j}^u &= u_{i,j} \\ V_{i,j} + n_{i,j}^v &= v_{i,j} \end{aligned} \tag{2.3}$$

where the indices (i,j) define the box considered, n_i^u and n_j^v represent the noise associated with the error of the measurement of the mean flow. The right hand-side of Eqs. 2.3 are the a priori values of the unknowns U_i and V_i .

An approximate balance for large scale steady flow is geostrophy on a β -plane, written as $\nabla U = -\beta V/f$, which ought to hold for the averaged velocities since structures smaller than the Rossby radius of deformation are presumably filtered out by the space-time averaging. A scaling of the standard error associated with the divergence gives $\delta_{div} \approx \frac{\delta U}{L} \approx \frac{10^{-2}}{10^5} = 10^{-7}$ where δU is the standard error associated with velocity and L is the size of the bin. A typical scaling of the divergence of geostrophic flow on the β -plane is $\beta V/f = \frac{10^{-11} \times 10^{-2}}{10^{-4}} = 10^{-9} \text{ s}^{-1}$. The expected size of the divergence of the flow is smaller than the error estimate by two orders of magnitude. We thus assume that the horizontal flow is non-divergent because the divergence of the flow is not dissociable from the noise.

The horizontal non-divergence constraint is written as:

$$(U_{i,j+1} - U_{i,j-1})(dy_{i+1,j} + dy_{i,j}) + (V_{i+1,j} - V_{i-1,j})(dx_{i,j} + dx_{i,j+1}) + n_{c.m.} = 0$$

Figure 2.3: Logarithm of potential vorticity ($f\partial\rho/\rho\partial z$). WOCE A24 Hydrographic section (june 1997) following the track Cape Farewell-Azores Plateau-Ireland. The black lines show the two neutral density layers $\gamma^n = 27.910$ and $\gamma^n = 27.975$ as given by the SAC climatology (continuous) and computed from the hydrographic section (dashed). The triangles denote the locations of the floats when crossing the section.

(2.4)

where $n_{c.m.}$ is the noise associated with this constraint, dx and dy are the bin steps which may vary next to the boundaries.

The third constraint we want to impose is the conservation of potential vorticity. Ideally, assuming no forcing and within uncertainties, the flow is constrained to follow lines of constant PV

Figure 2.4: Box used for the inverse model with a representation of the unknowns (uppercase) and the time-space average velocities (lowercase). Boxes are rectangular inside the basin (a) or fitted to the isobath (b) chosen as the closed boundary.

$$U_{i,j}q_x + V_{i,j}q_y + n_{p.v.} = 0 \quad (2.5)$$

where q_x and q_y are the eastward and northward components of the gradient of PV or q . Two expressions of q were used: f/H and f/h with H being the full water depth and h the thickness of the LSW layer (Fig. 2.3). Here, f is a function of latitude (and PV conservation gives rise to a divergence that is ignored since it is within the noise level). We define the core of LSW as the water contained within the two neutral density surfaces $\gamma^n = 27.910$ and $\gamma^n = 27.975$. These surfaces were chosen from the vertical extent of the LSW low potential-vorticity signature. This definition of the LSW boundaries is similar to the one proposed by Rhein et al. (2002) who used data obtained in 1997. The LSW thickness h was computed from the Hamburg Special Analysis Centre (SAC) climatology on a 0.5 degree resolution grid (Gouretski and Jancke, 1998).

Boundary formulation. At the closed boundaries, the boxes are fitted to a chosen isobath (1500m for the Labrador Sea Water, see Fig. 2.4). We have no measurement of velocity immediately along the boundary. The initial guess for the flow parallel to the boundary is not determined by averaging but by assuming that it is equal to the closest measurement, i.e. in the example of Fig. 2.4-b the V component is written

$$V_{i,j-1} + n_v = v_{i,j} \quad (2.6)$$

By introducing this formulation, it is possible to impose at a later step a constant boundary value for the streamfunction at the intersections of the 1500m isobath and the grid. Also, the introduction of artificial velocity vectors at the boundary insure that the boundary currents will be represented in the calculation. The average velocities at boundary gridpoints are usually highly significant; however, any smoothing implicit in the inverse model will include part of the weaker adjacent flow and bias the boundary value low.

In the area close to Flemish Cap the high eddy variability leads to high errors and the average float velocity does not show a definite southward flow. Mooring data, on the other hand, consistently show a significant average southward flow next to the boundary near 1500m depth (Schott et al., 2004). Clearly, time dependent fluctuations at various scales

(high frequency as well as eddies) expel the floats from the boundary region and result in a poor sampling there. The observed (Eulerian) boundary flow is the southward continuation of part of the Deep Western Boundary Current from the Labrador Sea (Fischer and Schott, 2002; Meinen and Watts, 2000). We introduce a southward velocity of 2 cm/s at the western boundary near 48°N, west of 42°W, to represent the mass sink (c. 3 Sv) on the LSW layer. This sink is the only constraint that we impose at the southern boundary.

Gauss-Markov solver. The linear system of equations is over-determined because the number of equations exceeds the number of unknowns. It has no exact solution apart from zero. However a solution can be obtained by minimizing the noise introduced in each equation (Wunsch, 1996). The measurements are noisy due to various factors including mesoscale variability, irregular spatial and temporal distribution of the floats and *array bias* introduced by float dispersion (Davis, 1991). Thus we cannot expect the above equations to be satisfied exactly. The unknowns were determined using the Gauss-Markov solver, which assumes prior knowledge of the error covariance matrix of the constraints and covariance matrix of the unknowns. Only the error variances are taken into account (the covariance matrix are diagonal) and were computed for each equation using the standard error on the mean velocities. The U and V were computed after writing the constraints into the matrix form

$$EX + noise = Y \quad (2.7)$$

where E is a design matrix, X is a single velocity vector made by combining both U and V, and Y a vector containing the right hand sides of Eqs. 2.3 to 2.6.

Errors estimates. The standard error associated with the time-space averaged velocities are usually large. In our analysis, only 48% of the velocities are significant at 95%. The most significant values are located in the boundary currents along Greenland and in the Irminger Current. Within the interior, most of the velocities are barely significant or similar in size to the error ellipses. In the Newfoundland Basin (east of Flemish Cap) larger errors arise from the large eddy variability of this area. Even though almost half of the time-space averaged velocities are statistically insignificant we include all values and standard errors in the calculation, retaining in that way as much information as possible.

It is known that LSW is subject to a significant decadal variability (Talley and McCartney, 1982) and therefore the PV distribution given by the climatology may not be representative of the particular circulation sampled by the floats. Fig. 2.3 shows a WOCE (A24, June 1997) section of the isopycnal potential vorticity across the area of study. It presents a comparison between the measured thickness of the LSW and the climatological one. Greatest differences are observed in the Irminger Basin on the western end of the section but the characteristic “dome” shape of the isopycnals in this area is clearly reproduced by the climatology. We further consider that it is of interest to use the climatological PV field as a constraint on the large scale mean flow, to help overcome biases in the (temporally and spatially distinct) datasets and to maintain consistency across the whole basin.

Streamfunction. In the previous section we have computed an adjusted velocity field. Now we want to obtain a streamfunction field in order to have a global view of the circulation throughout the basin. With similar motivation, previous studies (Lavender et al., 2000; Bower et al., 2002) used the objective analysis method introduced by Gandin (1965) and discussed by Bretherton et al. (1976) where the streamfunction is estimated directly from the velocities. The objective analysis is based on the Gauss-Markov theorem which assumes prior knowledge of the spatial covariance of the streamfunction and noise level of the measurement. The spatial covariance can be simply a Gaussian function of the distance between two measurements. Anisotropy can be introduced through explicit dependence of the spatial covariance on barotropic potential vorticity f/H (alternatively f alone in low latitudes, (e.g. Davis, 1998), or H alone at high latitudes) or through some intermediate version of this constraint reflecting an equivalent barotropic flow (e.g. Gille, 2003). Although the objective analysis method has been used repeatedly, especially for float velocity data it has the disadvantage of not taking into account any boundary condition. In our problem of deep circulation in the North Atlantic, boundary conditions matter because the basin is closed in the north by the topography (1500m isobath) and the data extend to the boundary. We chose to compute the streamfunction by solving the Poisson equation using the method described in the appendix. Straneo et al. (2003) also used a Poisson solver to obtain a non-divergent flow field from objectively analyzed velocities of 700 m float data in the northwestern Atlantic (see also Carter and Robinson, 1987). The advantage of this approach is that the constraint of the closed boundary, which was taken into account in the calculation of velocities, is

carried through to the streamfunction. A narrow feature like the Reykjanes Ridge and its effect on the circulation are in this manner better represented. The imposed conditions are zero value at the closed boundary and the streamfunction gradient along the line of the open boundary as given by the normal velocities. As a final check, velocity vectors are re-derived from the streamfunction and compared to the original averaged values.

CHAPTER 3

RESULTS

3.1 Subpolar gyre

Streamfunctions for the solutions with barotropic and isopycnal PV constraints are presented in Fig. 3.1 and the associated standard error in Fig. 3.3. There is a tendency for large-scale isopycnal PV to agree with barotropic PV (LaCasce and Speer, 1999) so we expected minor differences between the solutions with different PV constraints. In both cases large-scale flow is clearly similar, but details are different, especially in the southwestern Labrador Sea and the area southeast of Cape Farewell. The subpolar gyre is defined on its northern and western sides by the Eastern and Western Deep Greenland Currents followed by the Labrador Sea Deep Western Boundary Current, which flows south out of the domain off Newfoundland (as it is forced to do by the imposed southward velocity). In order to understand the effects of this outflow constraint, the same calculation was made without it. It appears that its main effect is local and essentially forces the southward flow from the Labrador Sea out of the domain instead of (implausibly) turning east to join the NAC. Increasing the outflow (not shown) beyond the strength of the resolved southward boundary current simply drives local recirculation.

On the eastern side of the ocean, the subpolar gyre is comprised of the deep extension of the North Atlantic Current followed by the Deep Irminger Current flowing along the western and eastern slopes of the Reykjanes Ridge. Inside the subpolar gyre three dominant cyclonic recirculations are apparent: an Irminger Basin gyre, a Labrador Sea gyre and an elongated gyre in the Iceland Basin up to the Mid-Atlantic Ridge (MAR). The third one is the largest and extends from 51°N to 60°N . This recirculation includes water flowing along the eastern flank of the Reykjanes Ridge toward the Deep NAC and may also be subdivided into

two smaller scale cyclonic gyres, a hint of which might be visible in the chlorofluorocarbon distribution (Rhein et al., 2002). Southeast of Cape Farewell (57°N , 41°W), a marginally significant anticyclone appears in the barotropic PV solution, while the isopycnal case shows weak flow bringing water from the Labrador Sea into the Irminger Sea.

The Gauss-Markov solution gives standard errors on the velocities. By taking into account finite difference errors in the calculation of the streamfunction and by adding the errors inherent in the method described in the appendix, we obtained an estimate of the total error on the streamfunction (Fig. 3.3). It reaches a local maximum of 0.8 cm in the Newfoundland Basin where the time-space average velocities are insignificant. Along the open southern boundary of our calculation, error increases rapidly southward to 2 cm from 48°N to 42°N . This high value is, by more than 50%, accounted for by the streamfunction calculation and reflects a mass closure imbalance incompatible with the no net meridional flow condition implicit in the streamfunction calculation (see appendix). North of 50°N , errors are on average larger east of the MAR (0.5-0.6 cm) than west of it (0.3-0.4 cm), except in the western half of the Irminger Basin reflecting a lack of data there (see Fig. 2.2).

The magnitude of the streamfunction difference between the two PV constrained maps (not shown) was computed and divided by the standard error. A significant difference, corresponding to a ratio larger than one, is located in the Labrador Sea where, in the case of the isopycnal PV constraint, the cyclonic recirculation in the central Labrador Sea is accentuated and consequently the Deep Western Boundary Current weakened by about 30% (see Fig. 3.1). This pronounced recirculation is a direct effect of the closed PV contours of the Labrador Sea.

To facilitate interpretation of the PV constraint, the result without PV conservation is also shown (Fig. 3.2a), which is closer to previous results. In addition, a solution has been computed from a subset of the averaged float data with high significance (Fig.3.2b), that is, with small standard error. The selection process for this subset was arbitrarily chosen as the norm of the averaged velocity being greater than twice the standard error along the major axis of its error ellipse. Only a small fraction (8%), for the most part on the boundaries, passed the selection process, but many major features are reproduced, showing the basic role played by boundary conditions and boundary currents.

3.2 Comparison with previous results and discussion

Lavender et al. (2003) computed the circulation of the Labrador Sea Water at 1500m using an objective analysis of the Palace float data used here. The spatial covariance function of their analysis is anisotropic, depends on the barotropic potential vorticity, and has a correlation length scale of a 150 km. Their map shows a representation of the subpolar gyre circulation similar to ours. A few differences are, however, notable. The previous map shows a clear eastward flow connecting the Labrador Sea to the Irminger Sea. A similar but weaker flow is visible in our space-time averaged velocity map (Fig. 2.2). This flow does not appear on the barotropic PV solution (Fig. 3.1) but does appear on the isopycnal PV solution. The barotropic PV inversion presents instead a weak anticyclonic recirculation. The isopycnal PV inversion attenuates considerably the eastward flow at the southern tip of Greenland, but does show a direct communication between the Labrador Sea and Irminger Sea (Fig. 3.4). At the southern edge of this eastward flow is a westward flowing prolongation of the deep Irminger Current. These differences are close to error bounds, but the evident communication and difficulty with PV conservation is consistent with local forcing driving the eastward flow.

Bower et al. (2002) present a circulation of the LSW obtained from the RAFOS and MARVOR floats used in this study. Their objective analysis uses a isotropic spatial covariance function and a correlation scale of 100 km. Other than presenting more spatial structure than our study, their map shows a broader southward flow in the eastern basin, apparently originating from the deep NAC current. Both the Bower et al. (2002) and Lavender et al. (2003) representations show an ill-defined split of the deep NAC at the Rockall Plateau with a branch flowing north in the Iceland Basin and a branch going south. We do not observe much of a split (Fig. 3.1) as most of the deep NAC flow continues northward. A weak southward recirculation occurs with flow along the eastern flank originating near (49°N, 23°W).

The circulation around the Reykjanes Ridge presents an asymmetry with respect to the ridge (noted by Bower et al., 2002). The southwestward flow on the eastern flank is more intense than the northeastward flow on western flank (Fig. 3.5). Mean velocities reach about 10 cm/s on the east side as opposed to 2 to 4 cm/s on the west side of the ridge, where eddy

kinetic energy levels are higher (Bower et al., 2002). Of the 7 streamlines east of the ridge, 4 turn into the Irminger Sea, 1 joins the Labrador Sea, and 2 leak across the MAR to feed the deep North Atlantic Current before it passes over the Charlie-Gibbs Fracture Zone (CGFZ).

Paillet et al. (1998) estimated transport of LSW in the eastern North Atlantic by using an inverse model based on hydrography and a set of various dynamical constraints (see also Paillet and Mercier, 1997). They define LSW with criteria on temperature and salinity which account for mixing as it spreads eastward. Their results were 6.3 ± 1.2 Sv flowing eastward across 35°W , between 44°N and 54°N . This flow split into a northward recirculation of 1.5 ± 1.4 Sv and a southward recirculation divided into 1.7 ± 1.1 Sv on the western side of the ridge and 3.1 ± 1.0 Sv on the eastern side. They noted the weakness of their northward flow when compared to previous results (Schmitz and McCartney, 1993) of 13 Sv. Their southward transport is comparable to the value derived below, but the structure of the flow is significantly different from that implied by the float data.

Precise estimates of transport cannot be given by the present analysis because it does not account for the vertical structure of the circulation. However, considering a LSW average layer thickness of roughly 1000m and the fact that 1cm in geostrophic pressure is equivalent to $1000 \text{ m}^2/\text{s}$ in transport per unit depth, we can attribute one contour interval of the streamfunction to a transport of 1 Sverdrup. From this, the eastward flow over the Charlie-Gibbs Fracture Zone is 6-7 Sv. As it reaches the Rockall Plateau, most of it recirculates into the Iceland basin, leaving only about 1-2 Sv flowing southward. The circulation in the eastern basin (east of the MAR) is characterized by a broad southward drift of about 3-4 Sv across 45°N , which is fed by a broad cross-ridge zonal flow (Fig. 3.1) south of the CGFZ.

The cross-ridge flow south of the CGFZ was expected to be concentrated at the major fracture zones on this part of the ridge but this is not evident in the solutions, perhaps because of lack of resolution in the data. Cross-ridge flow is generated by a relatively low number of floats but is nevertheless required to balance mass in the flow which intensifies to the south on the eastern flank of the MAR. This flow is reduced in the isopycnal PV case compared to the barotropic PV case. As shown in Fig. 3.6, the cross-ridge flow south of the CGFZ is zero within error, except near 41°N , where it is westward, and $47\text{-}48^\circ\text{N}$, where flow in both directions occurs. Also, eastward velocities over the CGFZ tend to be attenuated.

Figure 3.1: Two cases of streamfunction calculation where two potential vorticity constraints are imposed: (a) barotropic f/H , (b) isopycnal potential vorticity of the LSW layer. Streamfunction is a geostrophic pressure in cm. The contour interval is 1 cm.

Figure 3.2: Same as Fig. 3.1. (a) No PV^{20} constraint is imposed, (b) a subset (8%) of highly significant velocities is used to conduct the inversion, the conservation constraint of baroclinic PV of the LSW is imposed.

Figure 3.3: Standard error on ψ in cm estimated from the error given by the Gauss-Markov estimator.

Different averaging procedures could result in somewhat stronger flow over the CGFZ, and a stronger southward recirculation or branching from the deep NAC, but they are not justified by the resolution of the data.

A bulk horizontal recirculation time may be estimated as the ratio of the LSW volume to gyre transport. By this measure the subpolar gyre as a whole has a circulation timescale of about 10 yrs. The timescales for recirculation within basins or for exchange between basins depends on flow structure that depends, in turn, on the constraints used in the model. A quantitative assessment of the value of a constraint requires a model that includes tracers as well as velocity data. Qualitatively, the CFC concentrations (Rhein et al., 2002) south of Greenland, for example, suggest that the isopycnal PV constraint might be better, as a more direct exchange between the Labrador Sea and the Irminger Sea results, and the large-scale

pattern of high CFC extending east in two branches is more consistent with the circulation associated with this constraint.

Figure 3.4: Detail on the Irminger Sea showing the isopycnal PV streamfunction (contours, cm), the potential vorticity distribution (shading), the average float velocities (white arrows) and the velocities obtained by the inversion (black arrows).

Figure 3.5: Velocities with error bars on both sides of the Reykjanes ridge. Shown are velocities computed from the streamfunction (bold line) and from linearly interpolating the bin-averaged velocities and errors (stars).

Figure 3.6: Cross-ridge velocities along the Mid-Atlantic Ridge (MAR). Velocities are projected normal to the line following the Rift Valley (Fig. 1.1). Shown are velocities computed from the streamfunction (bold line) and from linearly interpolating the bin-averaged velocities and errors (stars).

APPENDIX

STREAMFUNCTION

Once a velocity field was obtained with the method described in section 2.4 we computed the nondivergent streamfunction ψ of the field using the procedure presented here. Unlike in the framework of section 2.4 where the measurement noise was estimated *a priori*, the computation of the streamfunction should introduce very few additional significant errors (Carter and Robinson, 1987). However, as explained below, possible noise had to be taken into account along the boundaries. The method involves solving the Poisson equation

$$\nabla^2 \psi = \vec{k} \cdot (\nabla \wedge \vec{u}) = V_x - U_y \quad (\text{A.1})$$

subject to boundary conditions which consist in imposing a zero value at the closed boundaries and the normal derivative given by the outward flow. Using finite differences, the set of equations was written in the matrix form

$$\mathbf{A}_1 \psi = \xi \quad (\text{A.2})$$

$$\mathbf{A}_2 \psi + n = Y_0 \quad (\text{A.3})$$

where $\mathbf{A}_1$ contains the finite differences expression of the Laplacian, $\mathbf{A}_2$ contains both boundary conditions, ξ is the curl of the velocity field, Y_0 contains the zero boundary value for ψ as well as the values of the normal gradient along the opened boundary. The vector n contains the residuals allowed along the boundaries. It is introduced to compensate the divergence of the velocity field (i.e. because the net meridional transport across the open boundary must vanish). This additional adjustment has the moderate disadvantage of increasing the streamfunction estimation errors along the open boundary. The set of

equations is solved using least squares method. The cost function to be minimized is, following Wunsch (1996) notations¹

$$J = (\mathbf{A}_1\psi - \xi)^T(\mathbf{A}_1\psi - \xi) + (\mathbf{A}_2\psi - Y_0)R_{nn}^{-1}(\mathbf{A}_2\psi - Y_0) + \psi^T S^{-1}\psi \quad (\text{A.4})$$

The solution for ψ is

$$\psi = (\mathbf{A}_1^T \mathbf{A}_1 + \mathbf{A}_2^T R_{nn}^{-1} \mathbf{A}_2 + S^{-1})^{-1}(\mathbf{A}_1^T \xi + \mathbf{A}_2^T R_{nn}^{-1} Y_0) \quad (\text{A.5})$$

where the weight matrix R_{nn} is diagonal and sets the expected size of the residuals allowed along the boundaries. The second weight matrix S is defined by

$$S^{-1} = \alpha^2 R_{\psi\psi}^{-1} \quad (\text{A.6})$$

where α^2 is a trade-off parameter and $R_{\psi\psi}$ is a normalized spatially Gaussian smoothing (weighting) function having a e-folding scale of 100 Km. The matrix S^{-1} was introduced for regularization purposes where $R_{\psi\psi}$ adds a smoothing condition and reflects our intention to obtain a smoothed streamfunction (Neumaier, 1998). Once ψ is computed, the residuals of the calculation are found to be smaller than the curl by a factor of about 10^{-4} , except for the boundary values which have the expected size set by the weight matrix R_{nn} .

After the computation, ψ is interpolated to a finer $0.25^\circ \times 0.25^\circ$ grid using a biharmonic spline interpolation scheme (Sandwell, 1987). Note that the interpolation does not affect the computed values of ψ on the $2^\circ \text{longitude} \times 1^\circ \text{latitude}$ grid (no additional errors are introduced by the interpolation).

¹ $J = n_1^T n_1 + n_2^T W^{-1} n_2 + \psi^T S^{-1} \psi$

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