Buoyant Eddies Entering the Labrador Sea Observed with Gliders and Altimetry

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ABSTRACT

Intense, buoyant anticyclonic eddies spawned from the west Greenland boundary current were observed with high-resolution autonomous Seaglider hydrography and satellite altimetry as they entered the Labrador Sea interior. Surveys of their internal structure establish the transport of both low-salinity water in the upper ocean and warm, saline Irminger water at depth. The observed eddies can contribute significantly to the rapid restratification of the Labrador Sea interior following wintertime deep convection. These eddies have saline cores between 200 and 1000 m, low-salinity cores above 200 m, and a velocity field that penetrates to at least 1000 m, with 0–1000-m average speeds exceeding 40 cm/s. Their trajectory, together with earlier estimates of the gyre circulation, suggests why the observed region of deep convection is so small and does not occur where wintertime cooling by the atmosphere is most intense. The cyclostrophic surface velocity field of the anticyclones from satellite altimetry matched well with in situ dynamic height baroclinic velocity calculations.

1. Introduction

Deep convection in the Labrador Sea forms Labrador Sea Water (LSW), a major contribution to the global ocean’s meridional overturning circulation (MOC). Wintertime cold-air outbreaks from the Canadian landmasses chill the Labrador Sea to increase the density of near-surface waters. Gravitational instability causes water parcels to sink as deep as 2000-m depth during harsh winters. The convected water mass, LSW, joins the deeper, denser overflow waters from farther north, yet its production rate is uncertain by nearly an order of magnitude (Bailey et al. 2005; Pickart and Spall 2007). Decadal temporal variation in the LSW formation rate seems to vary in phase with the strength and the extent of the subpolar gyre’s lateral circulation (Hakkinen and Rhines 2004; Lazier et al. 2002), which, in turn, influences the inflow of Atlantic water to the Nordic Seas (Hatun et al. 2005).

The intensity and depth of convection are primarily governed by heat loss to the atmosphere (Lazier et al. 2002), but the invasion of buoyant water from the surrounding shelves can modulate the convection (Bailey et al. 2005; Curry and McCartney 2001). Convection in the Labrador Sea is followed by restratification roughly from April to December when freshwater appears at near-surface depths and heat and salt converge at middepths. Salinity stratification near the surface layer and thermal stratification at middepths constitute the main resistance to convection. The rate of freshening is higher than can be explained by atmospheric precipitation minus evaporation (P − E) (Sathiyamoorthy and Moore 2002), so lateral exchange with the shelf regions is suspected to explain the seasonal restratification. Yet, the origin of the rapid invasion of freshwater into the surface layer has remained a puzzle. The regional source of the low-salinity water could involve the massive flow of low-salinity water from the Baffin/Hudson Bays (Houghton and Visbeck 2002; Khatiwala et al. 2002), but this freshwater is likely to be trapped along the Labrador continental shelf (Myers 2005). The West Greenland Current provides an alternate source of low-salinity near-surface water (Schmidt and Send 2007) and warm, saline middepth water as well. The observations discussed herein support this restratification pathway.
Recent observational and modeling studies have emphasized the importance of anticyclonic eddies ("Irminger rings") for the shelf–basin heat exchange in the Labrador Sea, but their contribution to the freshwater budget is still unknown (Eden and Boning 2002; Katsman et al. 2004; Lilly et al. 2003). Although the eddies are well documented from their surface signature, the only in situ observations of them near their generation site were made in summer 1997 by three profiling RAFOS floats that were captured within one anticyclone and two cyclones (Prater 2002). Prater’s study is a precursor to the work herein, both in profiling the water column in eddies involved in the separating boundary current and in applying satellite altimetry. Details about the size, strength, sense of rotation, and vertical water mass composition of such eddies as they are shed from the boundary current remain largely unknown. Lacking information on the initial composition of the anticyclones, Lilly et al. (2003) and Katsman et al. (2004) discuss only eddy heat fluxes and disregard the possibility that the eddies can have a fresh and cold top, and that they can transport salt at middepth.

The main impediment to understanding these eddies is a lack of observations during the convection period (January–March) resulting from the very harsh weather and the relatively small eddy sizes. We have surmounted these difficulties by using long-range autonomous underwater vehicles called Seagliders (Eriksen et al. 2001). These remotely controlled vehicles can carry out missions with lifetime exceeding six months, traveling typically 5000 km, even through harsh winter seas. They surface about three times per day, at which time course changes, sampling rates, and dive profiles can be reset. Hence, they are ideally equipped to sample complex eddy fields.

The goals of this study are first to describe the structure of these eddies, and second to estimate their potential (opposing) transports of freshwater at the near surface and middepths. Eddy heat transport is also estimated and will be compared with the literature. The main dataset comes from two Seagliders, while Argo floats, mapped satellite altimetry, and a sea surface temperature (SST) image are used to put the Seaglider data into a temporal and spatial context. Data from the Seagliders, satellite altimetry, and Argo floats are described in section 2. The large-scale oceanographic context and some initial indications of eddies are given in section 3. A detailed analysis of one particular eddy is presented in section 4, and its properties are given in section 5. An estimate of the shelf–basin eddy exchange of fresh–salt water and heat is given in section 6, followed by a summary in section 7. Some further considerations are presented in section 8.

2. Data

a. Seagliders

Seagliders are small (1.8-m hull), reusable, long-range, and buoyancy-driven autonomous underwater vehicles (AUVs) designed to glide from the ocean surface to as deep as 1-km depth and back while collecting profiles of temperature, salinity, dissolved oxygen, chlorophyll fluorescence, and optical backscatter. Details of their design are given in Eriksen et al. (2001). They are commanded remotely and report all of their measurements in near–real time via Iridium satellite data telemetry. They use GPS navigation at the sea surface to dead reckon toward commanded targets. The difference between dead-reckoned and actual displacements is used to estimate the depth-averaged horizontal velocity. By taking two GPS fixes typically 7–9 min apart, bracketing communications activity, an estimate of surface current is also obtained. Estimates of geostrophic or cyclostrophic shear can be integrated vertically and referenced to the depth-averaged current to obtain sections of absolute geostrophic current with typical errors of the order of 0.01 m s\(^{-1}\). The difference between observed and expected descent rate gives a profile of vertical velocity, showing internal wave and turbulent deep mixing activity.

The in situ data considered here were collected by a pair of Seagliders, designated sg014 and sg015, launched in Davis Strait from Research Vessel (R/V) Knorr in September 2004 (Fig. 1a). Toward the end of their 7-month deployment, they reached the region of strong eddies southwest of Greenland after crossing the central Labrador Sea from Hamilton Bank in late winter. Sg014 reached the seaward edge of the west Greenland continental shelf off Cape Desolation before traveling roughly parallel to the Greenland coast northward. Sg015 was guided northward from Hamilton Bank near 55°W. The intention was to recover both gliders at Fyllas Bank, offshore from Nuuk, Greenland. Sg015 expended its battery pack after more than 6 months at sea without reaching Fyllas Bank, having been blocked in the last few weeks of its mission by the eddy activity discussed herein, and was left adrift. Sg014 reached Fyllas Bank and was recovered after 7 months, 1 week at sea, completing the longest AUV mission to date. Its added endurance was enabled by having made fewer shallow dives than sg015 during its mission. Seaglider sg014 traveled a total of 5000 km over the ground, making 1248 profiles, and sg015 traveled 4500 km, with 1230 profiles.

Seagliders collect measurements along saw-tooth paths with a slope of 1:1–3.5 through the ocean (Eriks-
sen et al. 2001). Horizontal speeds through the water, averaged over a dive cycle, were 0.20–0.25 m s⁻¹. Dive and climb rates were maintained near 0.06 m s⁻¹, resulting in roughly 9-h dive cycles. Vertical resolution was set to be 0.3 (0–10), 0.6 (10–100), 1.2 (150–300), and 2.4 (>300) m. Seaglider salinities are based on samples recorded by an unpumped Seabird conductivity cell. A correction for thermal inertia errors is based on the flushing rate of the cell inferred from the glider forward speed, itself estimated from vehicle buoyancy and pitch angle. The salinities are accurate to about 0.01 in the upper water and 0.005 at depths deeper than the halocline. Pre- and postmission calibrations were only done for sg014 (since sg015 was lost) showing a sensor drift of only 0.002 in salinity. The average horizontal resolution of the Seagliders is approximately 2 times the dive depth over the glide slope, or 6 km for 1-km-deep dives in still water.

The midway horizontal position between the start and end of a dive cycle will be denoted as a dive position, and the depth-averaged horizontal velocity during the dive cycle will be associated with this position. Averaging the potential temperature and salinity (θ–S) profiles from the dive and the climb gives an average dive θ–S profile that likewise will be associated with the dive position.

b. Satellite altimetry

Both along-track and gridded data are used in order to give a detailed description of a specific eddy (along-track data) and a broader overview over the eddy field (gridded data). The along-track data are obtained from the Jason-1 sea surface height (SSH) anomaly product (information online at http://podaac.jpl.nasa.gov). There are 127 orbits per cycle and each orbit has a period of 112 min. The ground tracks of the orbits are about 100 km apart in the Labrador Sea and these are repeated almost exactly every 9.92 days. Along-track measurements are approximately 1 s apart, corresponding to a 6-km separation. The sea surface height anomaly represents the difference between the best estimate of the sea surface height and a mean sea surface as averaged over the period of 1992–2005. This has been corrected for atmospheric effects (ionosphere, wet and dry troposphere), electromagnetic bias, and other contributions [ocean tides, pole tide, and inverse barometer; see Picot et al. (2003)]. The mapped sea level anomaly (MSLA) data combine altimeter data
from the Ocean Topography Experiment (TOPEX)/Poseidon, Jason-1, Environmental Satellite (EnviSat), and Geosat and maps these onto a $\frac{1}{2}^\circ$ Mercator grid, using optimal interpolation while accounting for long-wavelength errors (Ducret et al. 2000). Two maps are produced every week. The physical corrections mentioned above are applied to the along-track data before gridding. Mapping errors are less than or about 20% of the signal variance (Le Traon and Ogor 1998). The gridded altimeter product was produced by the Collecte Localisation Satellites (CLS) Space Oceanography Division (information online at http://www.jason.ocean-obs.com).

c. Argo floats

Argo floats collect high-quality temperature and salinity profiles from the upper 2000 m roughly every 10 days along with currents from 1000-m depths. After these floats have spent typically 10 days drifting at 1000-m depths, they sink to 2000 m and then rise to the surface over a 6-h period, while measuring temperature and salinity. At the surface, satellites position the floats and data are transmitted. The floats then sink to 1000-m depth again and the cycle is repeated. Only delayed-mode data are used and the uncertainty in the salinity measurements is therefore below 0.01 psu. Data from these floats are publicly available from the International Argo Project, which is a pilot program of the Global Ocean Observing System (online at http://www.argo.ucsd.edu).

3. Three eddies—From the boundary currents to the interior basin

A sketch of the Labrador Sea boundary current system is shown in Fig. 1a. The northward flows along the west Greenland shelf consist of the three main components. The deep western boundary current (DWBC) is guided by the bottom slope cyclonically around the Labrador basin along the 3000-m isobath. The West Greenland Current (WGC) is confined mainly between the 1000- and the 2000-m isobaths. The more sluggish remains of Irminger Water (IW; $\theta \sim 3.4^\circ$C and $S \sim 34.95$, where $\theta$ is potential temperature referenced to the sea surface) are sandwiched between these two (Cuny et al. 2002) (Fig. 1a). The DWBC is weakly sheared and carries water from the Iceland–Scotland overflows ($\theta \sim 3^\circ$C; $S \sim 34.92$) and the Denmark Strait overflow ($\theta \sim 1.5^\circ$C; $S \sim 34.9$), collectively termed the Eastern Overflow Waters (EOW). The West Greenland Current is more baroclinic, with maximum currents and maximum vertical shear located near the 2000-m isobath (Cuny et al. 2002). This current lies on a very steep continental rise and is associated with a sharp shelfbreak front between the fresh and cold West Greenland Current Water (WGCW; $\theta \sim -1.8^\circ$C, $S \leq 34.5$), which originates in the Nordic seas, and the warmer and more saline underlying IW. Transport estimates for these flows are as follows: WGC, 3 Sv (1 Sv $= 10^6$ m$^3$ s$^{-1}$); IW flow, 11 Sv (Clarke 1984); and a combined flow of WGCW, IW, and EOW past Cape Farewell is between 34 and 50 Sv (Clarke 1984; Gans and Provost 1993; Reynaud et al. 1995). Speeds in the West Greenland Current weaken considerably near 61°N where a portion of this flow follows the deep flow to the southwest, while the remaining flow continues north as the West Greenland Current Extension (Cuny et al. 2002). These boundary currents encircle the large body of LSW ($\theta \sim 3.4^\circ$C; $S \sim 34.84$) that occupies most of the central Labrador basin.

Important geographical connections are seen (Fig. 1b) in a combined plot of late-winter surface velocity variance, wintertime convection depth, and Eulerian streamfunction. The maximum in surface current speed based on late-winter altimetric SSH anomalies fits within the cyclonic gyre determined by the Profiling Autonomous Lagrangian Circulation Explorer (PALACE) floats of Lavender et al. (2002). The maximum in wintertime convection depth for 1997 is isolated from this region, lying south in the region of Eulerian recirculation. Other ship-based and Seaglider sections verify the weakness of deep convection northwest of the mapped convection depth here. It thus appears that the region of deep convection coincides with recirculation regions that are “protected” from the west Greenland outflow, with its influx of buoyant stability. While there is significant year-to-year variability in deep convection, owning largely to air-sea heat flux variability and water column stability, the general picture in Fig. 1 is quite robust.

a. Indication of eddies

Embedded in the mean circulation are strong mesoscale eddies, likely associated with instability of the west Greenland boundary current. The high buoyancy within the Irminger rings and the resulting strong anticyclonic circulation around them is associated with knolls in the sea surface topography. The two Seagliders encountered three major upward SSH anomalies during February and March 2005. An overview is given in SSH maps (Fig. 2). The numbered positive SSH anomaly features (1–3) were generated in the 61.5°–62.0°N, 52°–53°W region, and propagated southwestward just north of the 3000-m isobath with drift speeds of roughly 0.15 m s$^{-1}$ (Fig. 3). About 16 March, sg014
Fig. 2. Three anticyclones (numbers) illustrated by their sea surface signature in the gridded altimetry data (MSLA): (a)–(l) 16 Feb–26 Mar. The positions of sg014 and sg015 during the time interval ±4 days relative to each sea surface height field are shown as white dots. The white dashed arrow in (a) shows the propagation track shown in Fig. 3, and the black dashed arrows show the trajectory of eddies 2 and 3.
became entangled in an eddy associated with SSH feature 3 just as it was forming near the Greenland shelf, while sg015 encountered feature 2 after having traversed much of the northern Labrador Sea (Fig. 2i). The SST image in Fig. 4, serendipitously obtained from relatively clear skies on 19 March when both gliders were near SSH maxima, shows a large, cold filament across the northern Labrador Sea, which seemingly originates from the Greenland shelf. Cold patches of water are associated with the two SSH anomalies, indicating that the features are associated with cold and (by inference) fresh caps.

As sg015 moved north from the interior Labrador Sea (Fig. 1a) it first encountered SSH feature 1 just south of 60°N and then became entangled in the eddy indicated by anomaly 2 at about 60.5°N (Figs. 2e,f). In the vicinity of the SSH maxima (especially 2) it sampled strong θ–S anomalies with cold and fresh water in the surface 200 m, and warmer and saltier water in the lower 200–1000 m (Figs. 5 and 6a). The hydrographic characteristics to the north of this transect resemble those of the boundary current system, while the water column is very homogeneous farther south, resulting from deep convection (Fig. 6a).

b. Boundary water influx

The hydrographic changes in the convection region (56°–58°N, 51°–55°W) during the winter of 2005 are described using all θ–S profiles sampled by Argo floats within this region. These are grouped into the months from December to March and plotted in a θ–S diagram (Fig. 6b). The numbers of profiles made during each month are 6, 8, 8, and 5 for December, January, February, and March, respectively. The θ–S characteristics of the upper 1200–1300 m converge toward the 2005 vintage of LSW (θ = 3.4°–3.5°C, S = 34.84–34.85) during the convection period from December to March. But already in March, two of the five profiles within the convection region (Fig. 1b) sampled intrusions of freshwater in the surface 100- and 150-m layers, respectively, reflecting a very sudden and strong restratification of the water column.

Interestingly, the upper-layer θ–S properties in anomaly 2 are similar to the properties of the newly formed fresh cap in the convection region in March (Fig. 6b), only with 0.75°C higher potential temperatures and slightly lower salinities. The National Centers for Environmental Prediction (NCEP)-derived heat loss in the convection region is on the order of 200–300 W m⁻² in February, and a 200-m-deep slab of water, isolated from lower layers, would take 20–30 days to cool 0.75°C under such atmospheric cooling. It is therefore possible that the intruding cap of fresher water in March stems from features such as anomaly 2.

The annual net input of freshwater into the Labrador Sea surface layer is 60 cm (Lazier et al. 2002). Using data from Ocean Weather Station Bravo (OWSB; 56.75°N, 52.5°W) (1964–74) and PALACE floats (1996–2000), Straneo (2006) finds a spring peak (April–June) to freshwater inputs (25 cm in the
0–200-m layer), and a spring peak heat input of 390 MJ m\(^{-2}\) to the 200–1300-m layer. The mean annual heat contribution to the lower layer from 1996 to 2000 was 0.9 ± 0.3 GJ m\(^{-2}\) (Straneo 2006). These climatological freshwater and heat input numbers will be used later, when the exchange efficiency of the eddies is assessed.

4. A Seaglider profiling an eddy

Important questions are whether or not the fresh surface layer (Figs. 5 and 6a) actually derives from the Greenland shelf, and whether such a transport could be due to one or several anticyclonic eddies. About 2 March, sg014 left the west Greenland (Fig. 2e) shelf near the region where the SST image showed a conduit of cold water 17 days later (Fig. 4). Although sg014 was gliding west during this time period, it was arrested near 62°N, 52°W, where it moved rather randomly (Figs. 2g,h). As it began to move west-southwest around 16 March (Fig. 2i), it was directed to glide to the north. Despite this, the glider continued to move west-southwest and followed the roughly cycloidal trajectory shown in Figs. 2j,k,l and 7. Sg014 closely followed SSH anomaly 3, which developed in a very similar way as that of anomaly 2 about a month earlier (cf. Figs. 2a,b and Figs. 2k,l). Sg014 therefore reveals key information about the characteristics of the anomalies and their influx of freshwater.
This gives an impression that $V^\text{max}$ is the maximum azimuthal velocity, which $V_\theta$ is the current ve-
tot, (Fig. 8a). If a northward-moving glider at some time is located within an anticyclonic eddy that is translating in a westward direction, it can be trapped and dragged along with the eddy. The trajectories outlined by the simulated trapped gliders are very similar to what was observed (Fig. 7), confirming that the feature observed by sg014 is probably an anticyclone. Navigation strategies can be developed based on this model.

### a. Cycloidal path

Sg014 made three complete loops along its cycloidal trajectory, each with a period of about 3.5 days (Fig. 7). The observed depth-averaged current velocities are largest and oriented toward the west-southwest, as the glider moves through the open part of each loop. As the glider loops around each little “eye,” the current velocities decrease and their orientations revolve anticy-

To illustrate the trapping mechanism, simulations were made where an eddy was modeled as having a solid-body rotating core out to a radius of $R_\text{e}$,

$$V_\varphi(r) = \frac{V^\text{max}}{R_\text{e}} r, \quad r \leq R_\text{e},$$

(4.1)

beyond which the azimuthal velocity decreased quadratically:

$$V_\varphi(r) = V^\text{max} \frac{R_\text{e}^2}{r^2}, \quad r \geq R_\text{e}.$$  

(4.2)

Here $V^\text{max}$ is the maximum azimuthal velocity, which occurs at the core radius $R_\text{e}$. The eddy was prescribed to drift at constant velocity $V_D$, so the total flow field is given as

$$V_\text{tot} = V_D + V_\varphi.$$  

(4.3)

where the vector $V_\varphi$ is tangential to the eddy center. Into this flow field, an array of model gliders headed due north with the velocity $V_\varphi$ was started at positions $(x_{g,0}, y_{g,0}) = (x_i, 0), \quad i = 1, \ldots, n$. The trajectory of each model glider was found by solving the differential equation

$$\frac{\partial}{\partial t}(\mathbf{x}) = V_\text{tot} + V_\varphi,$$

(4.4)

given that the eddy center begins translating from $\mathbf{x} = (x, y) = (0, 0)$ at time $t_0 = 0$. A 14-day-long simulation, corresponding to the period when sg014 was trapped by the eddy, was performed using values obtained from the presented dataset: $V^\text{max} = 0.5 \text{ m s}^{-1}$, $R_\text{e} = 21 \text{ km}$, $V_\varphi = 0.2 \text{ m s}^{-1}$, and $V_D = (U_D, V_D) = (-0.15, -0.05) \text{ m s}^{-1}$ (Fig. 8a). If a northward-moving glider at some time is located within an anticyclonic eddy that is translating in a westward direction, it can be trapped and dragged along with the eddy. The trajectories outlined by the simulated trapped gliders are very similar to what was observed (Fig. 7), confirming that the feature observed by sg014 is probably an anticyclone. Navigation strategies can be developed based on this model.

### b. Eddy center

The position of the eddy center at any given time is needed in order to portray its radial/vertical structure from the glider data. In the simulations, both glider positions and eddy center are known. The location of the eddy center can be found by low-pass filtering the glider positions with a box-car filter of a width equivalent to the eddy rotation period ($\sim 3$ days) and adding zonal and meridional offsets. The offsets are estimated as the mean difference between the filtered glider center and the eddy center (Fig. 8b). The error associated with the estimated offsets is of order a few hundred meters for the scales used in the simulation. Subtracting the location of the simulated eddy center from the simulated glider positions gives the glider positions and flow velocities in the reference frame of the eddy (Fig. 9b). The northward-moving glider is mainly bound to the southeastern quadrant of the eddy, where opposing current velocities with magnitudes up to more than twice the glider speed are found. This explains why the glider is translated west-southwestward.

Because both the observed glider trajectory and the observed depth-averaged velocities resemble the simulations (cf. Figs. 7 and 8b) it is reasonable to expect that the location of the center of the observed eddy can be estimated similarly from filtered glider positions. The offsets between the filtered track and the eddy center can be estimated from observed current vectors by assuming these are predominately azimuthal (Fig. 9a). The observed and the simulated currents and trajecto-

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path (Fig. 1) with translation speeds of 0.16–0.2 m s$^{-1}$. Larger errors are associated with the observed eddy-center locations than with those estimated from simulations because neither are observed eddy trajectories necessarily straight and steady nor are the eddies themselves truly circular and steady. We estimate the uncertainty of the radial position to be $\sim 1$–2 km.

c. Depth-averaged current cross section

Using estimated radial positions, a depth-averaged current cross section has been estimated (Fig. 10). This section is based on glider dives made during the period of the regular cycloidal motion (Fig. 7). All profiles made during this period are located between the eddy center and the radius of maximum current velocities, perhaps with an indication of decreasing velocities at the outermost profiles (Fig. 10). The eddy periphery was not well sampled during this period. Prior to this period, sg014 was headed westward and entered the eddy, where it was trapped as the heading was changed from west- to northbound. Five profiles were made as sg014 entered the eddy, but positioning these relative to the eddy center was less certain because the method

Fig. 8. Tracks of (a) simulated northward-heading gliders in a velocity field governed by a solid-body rotation eddy that is moving roughly west-southwest. (b) Details of the enhanced trapped glider in (a). The symbols correspond to Fig. 7.

Fig. 9. (a) Observed sg014 and (b) simulated glider in eddy-reference frames. A core radius of 21 km has been drawn for orientation.
used above was not feasible. By assuming that the azimuthal currents decrease quadratically away from the eddy, radial position can be estimated from (4.1) and the observed currents. Other formulations have been used to model velocity structure in eddies (e.g., exponential decay, etc.); however, the differences in overall structure are minor (Prater 2002). Reasonable choices for \( V_{max} \) and \( R_e \), are 0.5 m s\(^{-1}\) and 21 km, respectively, based on the observed velocity profile (Fig. 10). Because of the necessity of using a model to estimate the radial position of these glider dives, the outermost five points in the velocity profile are associated with larger uncertainties than the other points. A continuous depth-averaged current cross section was estimated by interpolating all the current–radial position pairs onto a 1-km grid and low-pass filtering them with a 3-km-wide box-car filter (solid curve, Fig. 10).

5. Structure of an Irminger ring

a. Potential temperature and salinity

Synthetic radial cross sections of potential temperature and salinity across the eddy were constructed by combining radial position estimates and depth profiles. Properties were interpolated onto a regular depth–radius grid with a 5-m vertical and 1-km horizontal separation. A triangular-based linear interpolation was used to construct the sections (Fig. 11) that were smoothed with a 3-km-wide box-car filter. The eddy was assumed to be circular and steady in making these maps from dives during the period when sg014 was trapped. The maps are most credible over radii from 3 to 28 km. The five profiles at a larger radius were located using the \( r^{-2} \) decay model, and hence are less reliable. Although this will influence the character of the eddy in the radius range of 28–35 km, it does not alter the main features.

The presence of a warm core from 300- to 800-m depth and a saline core from 300- to at least 1000-m depth are prominent middepth features in these sections. From its properties, the eddy core at middepth can be identified as IW carried from the relatively

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**Fig. 10.** The radial–azimuthal depth-averaged current cross section of eddy 3. The circles between 0 and 28 km are directly observed currents, whereas the circles beyond 28 km have been fitted to the \( r^{-2} \) decay shown with the dashed curve. The low-pass filter profile is indicated by the solid curve.

**Fig. 11.** (left) Average potential temperature (°C) and (right) salinity (psu) depth–radius sections of eddy 3. Solid vertical lines indicate average locations of sg014 dive cycles based on a constant offset between low-passed glider-track and eddy-center locations. Dashed vertical lines indicate dive locations that are based on a solid-body rotation model.
warm and saline Irminger Current. The near-surface waters, above 300-m depth, likewise are distinctly cold and fresh. These are WGCW, derived from the West Greenland Current.

b. Potential density and total velocity

The accompanying potential density anomaly (Fig. 12b) section illustrates the characteristic bowl shape of the eddy encountered by sg014. The total azimuthal current field within the eddy can be separated into a depth-averaged component $V_A$ and a sheared component $V_B$ whose depth average at each radial position vanishes:

$$V_{tot}(r, z) = V_A(r) + V_B(r, z).$$

The depth-averaged component is directly measured by the glider (Fig. 10), and the sheared component can be obtained from the potential density field and using cyclogeostrophy. Combining these (5.1), the total velocity field is obtained (Fig. 12a). Cyclogeostrophy in an anticyclone is a balance between a centrifugal term plus a pressure term, and the Coriolis term. The relative importance of the centrifugal term to the Coriolis term is given by the Rossby number ($Ro = U/L$), where $U$ and $L$ are the scales for the azimuthal velocity and the eddy radius. For eddy 3 the Rossby number is on the order of 0.2.

c. Sea surface velocity and height

Anticyclones appear as knolls in the sea surface topography, but the gridded altimetry product (Figs. 2–3) is too coarse to give a quantitative picture of the surface signature of individual eddies. Fortunately, the Jason altimeter satellite passed over eddy 3 on 2300 UTC 24 March, (Fig. 13), measuring its sea surface height cross section. A geostrophic surface velocity estimate was obtained from the observed sea surface by applying a moving box-car filter of ~12-km length (two altimetry points) (Fig. 13). From this the velocity profile of the

![Fig. 12](image)

**Fig. 12.** (left) Average total velocity (m s$^{-1}$) and (right) potential density anomaly relative to 500 m, $\sigma_{0.5}$ (kg m$^{-3}$). Vertical lines indicate glider dive locations as in Fig. 11.

![Fig. 13](image)

**Fig. 13.** An altimetry track crossing eddy 3 on 2300 UTC 24 Mar. The geostrophic surface currents, calculated from the observed SSH along the track (black dots), are shown with the associated arrows. The estimated eddy-center path is shown with crosses and the position when the satellite crossed the eddy has a circle in addition. The depth-averaged currents during two glider dives, before and after the altimeter measurements, are shown with two arrows. A circle with a radius of 21 km has been drawn for reference.
underlying anticyclone becomes clearly visible. The estimated path of the eddy center is shown with crosses in the figure and the estimated location of the core when the satellite passed over is illustrated with a circled cross. Depth-averaged current vectors as measured by the sg014 before and after the passage of the altimeter satellite (1900 UTC 24 March and 0330 UTC 25 March) are plotted for comparison. The directionality of the surface and the depth-integrated currents agree, but the depth-averaged currents are weaker than the altimetry-derived surface currents because the flow field around the eddy is surface intensified (Fig. 12a). This gives an independent verification that the estimated offsets from the low-passed glider track to the eddy center are reasonably estimated. It also shows that the altimetry swath passed very close to the eddy core. This explains why eddy 3 appears so clearly in the gridded SSH fields during period around 24 March (Figs. 2 k,l).

The satellite passed from southwest to northeast and the unfiltered SSH knoll is plotted in Fig. 14 (solid curve). The glider-derived sea surface profile is obtained by integrating the cyclogeostrophic balance and using the sea surface velocities from Fig. 12a. The agreement of the altimetry with the glider-derived SSH cross section is reasonably good, although the glider-derived cross section is somewhat broader and lower.

d. Comparison with earlier observations

The only other in situ measurements of an anticyclone during its formation were made by a profiling RAFOS float (Rossby et al. 1986), deployed at 61.23°N, 50.38°W in June 1997 (Prater 2002). The azimuthal velocity profile measured at 375-m depths revealed a core radius \( (R_e) \) of 25 km with maximum velocities \( (V_{max}) \) of 0.42 m s\(^{-1}\), which is very similar to the depth-averaged velocity profile of eddy 3 (Fig. 10). Eleven Irminger rings passing two moorings near OWSB during 1994–99 were described by Lilly et al. (2003). These had azimuthal velocities of 0.3–0.8 m s\(^{-1}\) at the surface and velocities of 0.2–0.4 m s\(^{-1}\) at 1000-m depths (their Fig. 18b), which is also comparable to what is found from eddy 3 (Fig. 12a).

Eddy 3 shows a characteristic structure with weak vertical shear from the center to core radius \( R_e \), strong, deep (1000 m) currents, and a very strong surface current near \( R_e \); and vanishing deep currents, but a strong surface current extending beyond \( R_e \) (Fig. 12a). A very similar velocity structure was measured at the moorings, as an Irminger ring passed by (Lilly et al. 2003, their Fig. 16d).

As the RAFOs float described by Prater (2002) was shed from the Greenland shelf, it measured cooler waters (3.9°C) overlying warmer waters (4.9°C) at float level, but slightly off the shelf the surface layer temperatures changed to 4.7°C–4.8°C. The Irminger rings reported by Lilly et al. (2003) had “domes” of warm, salty surface-trapped water, and this is in contrast to the very cold and fresh caps on eddies 2 and 3 reported here (see below).

During the period discussed here, the large anticyclones drift roughly with the background flow described by Lavender et al. (2002) (Figs. 1, 2, and 3). The mean flow at 700 m diverges around 58°–60°N, 55°–57°W, where some streamlines retroreflect to the east and into the central Labrador Sea, while some continue south along isobaths (Fig. 1). Both simulations (Eden and Boning 2002) and the MSLA data indicate that many eddies follow the Lavender streamfunction into the central Labrador Sea, while some follow the flow south along the western margin of the basin. Other evidence for the coincidence between eddy drift and background flow has recently been presented by Fu (2006).

This is not immediately consistent with the findings by Lilly et al. (2003) who present a southward eddy propagation of ~5 cm s\(^{-1}\), obtained from a southward expansion of the eddy kinetic energy (EKE) fields from the eddy generation area (Fig. 1a) during winter. The inconsistency could be explained by the fact that a southward-expanding EKE field gives a Eulerian view of eddy propagation while the present analysis shows the drift of individual eddies (Lagrangian view).

Questions about where and how these eddies eventually disintegrate are complicated and beyond the scope of the present work. They are, however, important in order to assess the restratification efficiency of
these eddies, and thereby understand their influence on deep-water convection in the Labrador Sea.

6. Revised flux estimates

Acknowledging the uncertainty of the eventual fate of the Irminger rings, we will estimate the number of eddies required to account for the spring restratification, given that they actually disintegrate in a recently convected region.

a. Freshwater and heat contribution from a single eddy

The freshwater and heat contributions of eddy 3 have been estimated for its surface layer and its lower layer. For a radially symmetric eddy, the salinity and potential temperature averaged horizontally over the eddy are given as

$$S_e, \theta_e = \frac{1}{r_e^2} \int_0^{r_e} 2\pi r(S, \theta)(z, r) \, dr,$$  \hspace{0.5cm} (6.1)

where \((S, \theta)\) and \((z, r)\) are the fields shown in Fig. 11, and \(r_e\) is the radius within which significant \(\theta\) and \(S\) anomalies are observed. This eddy radius is about 35 km for eddy 3 (Fig. 11). If an eddy depth layer of thickness \(D\), carrying a volume \(V_e\) of water with the salinity \(S_e\), was mixed horizontally into a homogeneous basin of LSW (interior Labrador Sea) while a corresponding volume \(V_{LSW}\) of LSW with salinity \(S_{LSW}\) were ejected from the basin, the net salt contribution to the basin would be

$$\Delta M_{salt} = \rho_e V_e S_e - \rho_{LSW} V_e S_{LSW},$$  \hspace{0.5cm} (6.2)

where \(\rho_e\) and \(\rho_{LSW}\) denote eddy and basin interior density, respectively. The volume of freshwater \(V_F\) needed to give the same salt exchange is

$$V_F = \frac{\rho_{LSW} V_e S_{LSW} - \rho_e V_e S_e}{\rho_{LSW} S_{LSW}}.$$  \hspace{0.5cm} (6.3)

By assuming that the convective region of the Labrador Sea is a circular basin of radius \(r_{LS} = 300\) km (Lilly et al. 2003), and neglecting the influence of density changes, the freshwater contribution per unit area from each layer of the eddy is given as

$$\Delta F = \frac{V_F}{A_{LS}} = \frac{(S_{LSW} - \bar{S}_e)}{S_{LSW}} \left( \frac{r_e}{r_{LS}} \right)^2 D,$$  \hspace{0.5cm} (6.4)

where the overbar denotes a depth average over the layer considered and \(A_{LS}\) is the area of the convection region. Similarly, the heat contribution per area for each layer is given by

$$\Delta H = (\bar{\theta}_e - \bar{\theta}_{LSW}) \rho_0 c_p \left( \frac{r_e}{r_{LS}} \right)^2 D,$$  \hspace{0.5cm} (6.5)

where \(\rho_0\) is a reference density and \(c_p\) is the heat capacity of seawater.

The freshwater and heat contribution per unit area and depth from eddy 3 to a convected Labrador Sea is shown in Fig. 15. It appears that a layer partitioning based on salinity alone should be at about 350 m, while a partitioning based on potential temperature should be at about 150 m. To show the sensitivity on the choice of interface, the upper- and lower-layer freshwater and heat contributions are calculated by partitioning at 200 and at 300 m, respectively (Table 1). These values are obtained by using \(r_e = 35\) km, \(\rho_0 = 1027\) kg m\(^{-3}\), \(c_p = 4000\) J (kg °C\(^{-1}\)), and the present year’s characteristics of LSW as observed by sgo15 (section 3b). Despite the 50% difference in choice of upper-layer thickness, the crude approximation into layers demonstrates that the upper layer of eddy 3, if mixed into the assumed con-
and is thus slightly wider than that measured by the altimeter (Fig. 13). An eddy radius of 35 km could be an overestimation, but θ and S anomalies certainly extend far beyond the core radius of 20 km, as estimated from altimetry. If the eddy thermohaline anomaly radius were reduced to 30 km, the required number of eddies (equivalent to 3) to account for seasonal and annual changes in Labrador Sea stratification would be 12 and 31 for the spring and annual freshwater additions and 12 and 28 for the corresponding heat additions. The basic message that only a fairly few eddies are required to account for the observed restratification is unchanged by this uncertainty.

d. Spring peak eddy fluxes

With the eddy climatology (1993–2000) of the Labrador Sea produced by Lilly et al. (2003), the freshwater and heat contribution estimates from eddy 3 (Table 1) can be put into context. The climatological eddy sizes are described by the core radius $R_s$, and the eddy strengths are described by the peak-to-trough SSH anomaly amplitude $\delta_s$. With a core radius of 20 km and an SSH anomaly $\delta_s$ in excess of 20 cm (Fig. 14), eddy 3 would be categorized as moderate in size, but strong. Reassured by the fact that eddy 3 is similar to many eddies reported by Lilly et al. (2003) and to the anticyclone described by Prater (2002), in the following we will assume that eddy 3 represents a typical large anticyclone and examine the transports only by large anticyclones.

The eddy-shedding process is highly seasonal with most eddy activity during the winter months of December–March (Lilly et al. 2003). Assuming that eddies shed between January and May contribute to the spring peak, there would be 5 of the very strongest anticyclones ($\delta_s > 20$ cm) and 10 with strengths above average ($\delta_s > 15$ cm) available, on average, according to the altimetry climatology. This eddy count represents all eddies appearing in a box in the northern Labrador Sea covering the eddy generation region. Six distinct anticyclones were shed and transported toward the Labrador Sea between January and March 2005 as revealed by the MSLA data. It appears likely that only strong eddies appear clearly in these fields.

Unfortunately, the Seagliders did not continue to sample eddies as they disintegrated or entered the central Labrador Sea. If the anticyclones actually do dissolve in the basin without much vertical mixing, 50% of the freshwater input to the surface layer and more than 50% of the salt and the heat convergence in the lower layer during the spring peak could be accounted for by the largest anticyclones alone. Considering that θ and S anomalies also can be carried by a larger number of smaller anticyclones, it seems likely that eddies can ac-

### Table 1. Heat and freshwater contribution from eddy 3 to a homogeneous basin containing LSW.

<table>
<thead>
<tr>
<th>Interface depth (m)</th>
<th>Freshwater (cm)</th>
<th>Freshwater (cm)</th>
<th>Heat (MJ m$^{-2}$)</th>
<th>Heat (MJ m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(upper layer)</td>
<td>(lower layer)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>200</td>
<td>2.5</td>
<td>-1.0</td>
<td>-2.5</td>
<td>43</td>
</tr>
<tr>
<td>300</td>
<td>3.0</td>
<td>-1.5</td>
<td>0.8</td>
<td>40</td>
</tr>
</tbody>
</table>

vection region, carries the equivalent of a couple of centimeters of freshwater averaged over the basin contributed from the west Greenland shelf, while the lower layer offsets that contribution by about one-half from ISW. In contrast, the heat transported by this eddy is equivalent to about 40 MJ m$^{-2}$, which is almost entirely due to the lower-layer (ISW) contribution.

b. Number of eddies required

The entire spring freshwater peak described by Straneo (2006) (section 3b) could be accounted for by 10 eddies equivalent to eddy 3 (see Table 1), and the annual net input of freshwater of 63 cm would require 25 eddies. An annual influx of 25 eddies would also export 25 cm of freshwater from the lower layer (add salt), which is comparable to the annual 21 cm estimated by Straneo (2006). The convergence of heat in the lower layer during the spring peak could be accounted for by 9 eddies and the annual contribution of heat would require 21 eddies.

c. Uncertainties resulting from eddy size

The estimates are quite sensitive to the eddy radius. Using satellite along-track data Lilly et al. (2003) found an average eddy-core radius $R_s$ of 23 km, but used an average radius of 20 km in their heat transport calculation and found that the eddies could not account for the annual heat transport into the Labrador Sea. They did, however, acknowledge that the estimates would double if an eddy radius of 30 km was used instead. By using an average radius of 25 km instead, Katsman et al. (2004) concluded that eddies could provide most of the heat transport (though they did not account for their cold tops).

From the altimetry track that crossed eddy 3 near its core, we find that $R_s \sim 20$ km (Fig. 13), but the θ and S anomalies related to the eddy extend about 35 km from the center (Fig. 11). It is therefore plausible that the studies cited above have significantly underestimated the average eddy radius.

The glider-derived core radius is 24–25 km (Fig. 10), and is thus slightly wider than that measured by the
count for at least 50% of the spring peak boundary–basin exchanges of temperature and salinity. Similar results on the heat transports have been obtained (Katsman et al. 2004; Lilly et al. 2003), but it has not been possible to estimate the eddy-induced freshwater transports before. A recent model study shows a clear restratifying effect of Irminger rings (Chanut et al. 2006, manuscript submitted to J. Phys. Oceanogr.), which, however, is confined north of 58°N. The present observations are not in conflict with these findings.

e. Uncertainties

The estimates of the eddy transports must be considered as crude because of many uncertainties. After all, the estimates offered here are based largely on a sample of a single eddy, albeit one whose internal structure was sampled.

The vertical division between the surface and the lower layer was chosen to be 200 m in order to compare with the transport estimates from Straneo (2006), but choosing a division at 300 m instead (Houghton and Visbeck 2002) would give larger freshwater transport and smaller heat transports (Table 1). The lower layer in (Straneo 2006) was defined as being 200–1300 m, while we use 200–1000 m because of the dive depths of the Seagliders. A thinner lower layer will contain less salt and heat than a thicker layer, and our lower-layer estimate is therefore an underestimate compared to that of Straneo (2006).

Different areas representing a convected Labrador Sea have been chosen in the literature (Katsman et al. 2004; Lilly et al. 2003; Schmidt and Send 2007; Straneo 2006). We have chosen to describe this as a circle with a radius of 300 km, which is roughly equivalent to the area within the 3000-m isobath (Lilly et al. 2003). According to Pickart et al. (2002) this choice is probably too large (Fig. 1b), meaning that the eddies are spread over too large an area, and the freshwater and heat estimates per unit area will therefore be underestimated. If the convection area, on the other hand, is much smaller than assumed, it is likely that many eddies will disintegrate outside this convection region, and thereby contribute little to the restratification there. As Fig. 1b emphasizes, the key to this problem lies in several geographical distributions—surface layer buoyancy, atmospheric forcing, gyres and recirculations, and topographic provinces. The disagreement between observations from OWSB and the PALACE floats (Straneo 2006) demonstrates that spatial inhomogeneities within the central Labrador Sea must be considered in budget estimates.

The freshwater and heat contribution by eddy 3 was calculated relative to the homogeneous water column after convection [(6.4) and (6.5)]. This estimate applies immediately after deep-water formation if the eddy dissolves in a convected region, but the contribution from each eddy will diminish during summer as the boundary–basin gradients become smaller.

Large interannual changes are observed in eddy kinetic energy (Lilly et al. 2003), so comparing the period of eddy count (1993–2000) with the period of PALACE floats (1996–2000) could introduce errors. The properties of the background stratification and the boundary current waters also change interannually, and this will influence the eddy transports.

The size distribution of the Labrador Sea eddies was discussed extensively by Lilly et al. (2003), but the linkage to the subsurface θ and S anomalies and their radial extent remains to be credibly founded.

The fresh WGC water is probably confined to the near-slope region (Fig. 4) during winter when the eddy shedding is taking place. Eddies formed near the shelf might therefore be more efficient in drawing freshwater off the shelf than are eddies formed farther offshore. It is therefore not certain that all eddies observed by Lilly et al. (2003) actually contain a fresh and cold top. This could explain the discrepancy between the eddies observed at the moorings (Lilly et al. 2003) and those presented here.

In addition, transport is shared by eddies and the background gyre (as well as an Ekman component in the surface layer). Estimates of the total offshore transport must await a better understanding of the background circulation, as might be gained by either regular glider deployments or moored instrumentation.

7. Summary

Resistance to deep convection provided by buoyant waters inflowing from the west Greenland shelf and boundary current shapes both the deep convection geography of the Labrador Sea and the rates of production of LSW. Seaglider data together with satellite altimetry data sampled in the winter of 2005 have made it possible to describe the composition of a typical strong “Irminger ring” and to estimate its potential influence on the early restratification of the Labrador Sea. The anticyclone consisted of cold, fresh West Greenland Current water in near-surface (0 to ~200 m) depths and warm, saline Irminger Sea Water at middepths (~200–1000 m). The largest azimuthal currents are observed about 25 km from its center, with 0.7 m s⁻¹ at the surface and 0.5 m s⁻¹ depth averaged over the top 1 km. It was weakly sheared to a radius of 25 km and was more
baroclinic out to its perimeter at about 35-km radius. Its Rossby number is on the order of 0.2. The sea surface height anomaly inferred from the subsurface Seaglider observations and the satellite altimeter-estimated sea surface height are in modest agreement; both showing a ∼0.2-m-high anomaly above the eddy.

Seaglider sg014 sampled the eddy from its formation region near 62°N, 52°W and as it propagated west-southwest on the northern side of the 3000-m isobath with velocities between 0.16 and 0.2 m s⁻¹. The glider outlined a characteristic cycloidal track within the eddy, which also was reproduced with a simple simulation. Mapped sea level data show that many eddies follow the same path and that translation velocities of about 0.15 m s⁻¹ are typical. Another Seaglider encountered two other eddies of the same character that had been translated along the same path and were found immediately north of recently convected water in the Labrador Sea. If mixed laterally in the central Labrador Sea after convection, an eddy would add about 2.5 cm of freshwater to the surface 200 m in depth, and would extract 1 cm of freshwater and add about 40 MJ m⁻² to the ~200–1000-m-depth range. Given the typical number of eddies shed each year, and that they mix laterally into the central basin, these eddies could account for at least 50% of both the freshwater input to the near surface and the freshwater loss and heat convergence at depths between ~200 and 1000 m during spring restratification.

8. Further considerations

Generation

It is possible that the constriction of the Greenland slope topography between 59.5° and 61.5°N intensifies the WGC, causing it to become unstable and shed eddies (Eden and Boning 2002). An interesting notion is that the satellite SST image presented by Prater (2002), the float captured by an anticyclone presented in the same paper, and the simulated shedding of anticyclones shown in Eden and Boning (2002) and Chanut et al. (2006, manuscript submitted to J. Phys. Oceanogr.) all indicated that shelfwater is ejected offshore very locally at 61.5°–7°N, 51°W.

The SST image in Fig. 4 shows a narrow conduit at the same location carrying cold (and by implication fresh) water from the Greenland shelf to a large cold pool farther offshore. This supports the notion of a surprisingly local shedding of anticyclones. The shallower isobaths that presumably steer the fastest portion of the WGC turn sharply eastward into what seems to be a large canyon just north of 61.5°N (Figs. 1a and 4), which could trigger the instabilities.

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REFERENCES


Khatiwala, S., P. Schlosser, and M. Visbeck, 2002: Rates and


