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On the spatial structure and temporal variability of poleward transport between Scotland and Greenland

L. Chafik1, T. Rossby2, and C. Schrum3

1Department of Meteorology/Physical Oceanography, Stockholm University, Stockholm, Sweden, 2Graduate School of Oceanography, University of Rhode Island, Kingston, Rhode Island, USA, 3Geophysical Institute, University of Bergen, Bergen, Norway

Abstract The flow north of warm subtropical water though the northeastern Atlantic is known to have many pathways that vary over time. Here we use a combination of upper ocean current measurements between Greenland and Scotland near 60°N and satellite altimetry to examine the space-time variability of poleward transport. The high-resolution scans of currents in the top 400 m show that the Reykjanes Ridge serves as a very effective separator of flow toward the Nordic and Labrador Seas, respectively. Whereas the Labrador Sea branch exhibits two mean flows to the north on the western slope of the Reykjanes Ridge, the eastern branch flows north in roughly equal amounts over the deep Maury channel and east of Hatton Bank including the Slope Current. There is also a well-defined southward flow along the eastern slope of the Reykjanes Ridge. The satellite altimetric sea surface height (SSH) data show good overall agreement with geostrophically determined -level difference from the repeat ADCP sections (1999-2002), but are unable to resolve the fine structure of the topographically defined mean circulation. The altimetric data show that variations in poleward flow west and east of the Reykjanes Ridge are strongly anticorrelated. They further reveal that the two eastern subbranches also exhibit anticorrelated variability, but offset in time with respect to the Labrador Sea branch. Remarkably, all these variations cancel out for the entire Greenland-Scotland section leaving a gradual decrease in sea-level difference of about 0.06 m over the 1993 to the end of 2010 observation period.

1. Introduction

The thermohaline circulation of the northern North Atlantic Ocean comprises two branches or components [Brambilla and Talley, 2008], one that leads to the Nordic Seas with its production of very dense water that overflows into the deep North Atlantic, and one that leads to the Labrador/Irminger Sea and its production of intermediate depth Labrador Sea water. Since both seas are fed by warm water from the Gulf Stream/ North Atlantic Current, also known as the upper limb of the meridional overturning circulation, it is of considerable interest to know the strength of this transport, and where and how this supply toward the two seas is partitioned and varies with time [Bersch, 2002; Håkkinen and Rhines, 2004, 2009; Rhein et al., 2011; Saran-afanov et al., 2012]. The last reference, which includes a comprehensive review of the literature, uses a combination of hydrography and altimetry to estimate meridional transport in three top-to-bottom density classes at 59.5°N. They note that a major challenge to their work is the limited number of sections with which to filter out the energetic mesoscale eddy field in the ocean. These reservations aside, they estimate the total meridional overturning circulation, meaning the net flow north between Cape Farewell and Scotland in the shallowest density class (\(\sigma_t \leq 27.55\)) to be 16.5 Sv, of which 15.5 Sv flows north east of the Reykjanes Ridge (RR). The corresponding return flows in the densest class (\(\sigma_t \geq 27.8\)) east and west of the RR water mass are 3.7 and 9.6 Sv, respectively. The shallow density surface is such that it never outcrops east of the RR, which means that waters less dense than this can cool and densify in the Iceland Basin, but not so much as to cross this isopycnal surface. Hence, water in this shallowest category must either flow into the Nordic Seas or cross the RR farther north into the Irminger Sea. Water in the densest class has its origins in the Nordic Seas, but as it overflows it entrains shallower water thereby both losing in density and increasing in transport. The branch flowing toward the Nordic Seas can be monitored quite effectively where it passes over the Greenland-Iceland-Faroes-Scotland ridge [Hansen and Østerhus, 2000; Jónsson and Valdimarsson, 2005; Hansen et al., 2008; Rossby and Flagg, 2012]. These openings make it easier to reduce uncertainties in
transport measurement. The most recent of these studies, based on ADCP-measurement of currents from the Norsø, a high-seas ferry in weekly service between the Faroes, Denmark and Iceland, reports a ~10% higher inflow than earlier estimates at 4.1 and 4.4 Sv transport through the Shetland Channel and between the Faroes and Iceland, respectively. These cross-ridge studies provide an effective integral measurement on the interannual and long-term variability of the meridional overturning circulation (MOC) warm-to-cold water transformation at high latitudes. Scanning currents accurately and at high resolution from a vessel in regular traffic gives us considerable insight in the spatial structure of currents, and the degrees of freedom for averaging accrue as the program continues.

During the period 1999–2002, a 150 kHz ADCP was operated on the Royal Arctic Line vessel Nuka Arctica, which operates on a 3 week schedule between Greenland and Denmark. This ADCP performed extremely well delivering over 60 ADCP sections between Cape Farewell and Scotland, eight of which via Iceland. This paper follows an earlier study by Knutsen et al. [2005], hereafter abbreviated as K-etal. Based on analysis of only 33 of the sections, K-etal noted that currents in Irminger Sea and Iceland Basin were strikingly oriented in parallel to RR. They also noted an asymmetry of eddy kinetic energy over the RR, high on the western side, and very low on the eastern slope. Here we use the ADCP data to estimate poleward transport between Greenland and Scotland. Due to the presence of drift ice in the East Greenland Current, the Nuka Arctica rounds the southern tip of Greenland at various distances from Cape Farewell making it impossible to include the East Greenland Current systematically, so it will be excluded from further consideration here. (But the Nuka Arctica ADCP data archive includes a lot data from this region as well as along the west Greenland coast up to 68°40’N.) The ADCP velocity data will be used two ways: (1) surface velocity can be integrated to obtain geostrophic sea-level difference across the Irminger Sea and Iceland Basin, which can then be compared with altimetric estimates of sea-level difference, and (2) the velocity profiles to 400 m depth can be bin-averaged across the ocean and integrated to yield total upper ocean transport.

A detailed inter-comparison between ADCP and altimetric estimates of sea-level difference along the Oland line between New Jersey and Bermuda was recently completed by Worst et al. [2013], who noted close agreement out to 300–500 km distances. We find good overall agreement here as well, and use this to estimate upper ocean transport variability along the Nuka Arctica line since the start of satellite altimetry. The paper is organized as follows: the next section gives a brief summary of methods and data of both ADCP and altimetric methods since these are well known from other studies. Section 3 has two parts, the first focuses on the ADCP findings, and the second on the corresponding altimeter results. This is followed by a discussion in section 4. A brief summary ends the paper.

2. Methods and Data

2.1. ADCP Operation

The ADCP works by repeatedly transmitting acoustic pulses of a certain frequency in four oblique directions from beneath the ship. Immediately after each transmission the instrument switches to its listen-mode to measure the Doppler shift of the backscattered signal as a function of time (= increasing depth). This way one obtains a profile of relative motion between the ship and the water column along each of the acoustic beams. Next, on the assumption that currents at each depth are horizontal and nondivergent, the along-beam velocity components are remapped into horizontal and vertical components relative to the ship. Last, by subtracting out the ship’s movement and heading as determined from GPS navigation and a GPS-based compass, one obtains the absolute ocean current velocity (Flagg et al. [1998], provide a detailed description of a typical installation). The accuracy of water velocity depends directly upon accurate ship speed for the along-track direction and ship heading for the cross-track component. Keeping the heading error to a minimum is crucial since at 16 knot vessel speed a 1° compass error with result in a cross-track velocity of 16 × 1.852 × sine (1°) = 0.14 m s⁻¹. Fortunately, the heading error can be determined directly in shallow water by comparing the integral of vessel movement over the bottom (with the ADCP in bottom-track mode) with the GPS-determined displacement. This calibration is done for each transit, the RMS scatter of which is 0.07°. See Appendix A for a detailed discussion of velocity and transport uncertainties from the ADCP.

The ADCP database includes data from some 66 crossings of various completeness between Greenland and Scotland (cf. Figure 1). The zonal lines along 59°–60°N indicate the eastbound transits at a nearly east constant heading toward the Scotland shelf break. The curved lines are westbound transits along great circle
arcs toward Cape Farewell. The routing reflects an effort to minimize impact of weather. Infrequently (8 transits) the Nuka Arctica also stops in Reykjavik. Ten transits go through Prince Christian Sound, about 50 km up the coast from Cape Farewell. As the figure shows there are scattered routes outside these envelopes; almost certainly these are due to heavy weather. Despite the often-inclement weather in the northern Northeast Atlantic, the quality of the ADCP data is quite high. This appears to result from a combination of two factors: the deep 8 m draft of the vessel, and the deep and small bow thruster opening that minimizes the down-scooping of air when the bow breaches the surface in severe weather. As a result, the 150 kHz ADCP often returns velocity data from as deep as 400 m. The velocity data used here have been detided by computing the spatial dependence of the principal tidal components (M2, N2, S2, L1, and O1) in the upper 100 m using a least square method \cite{Dunn, 2002; Wang et al., 2004; Rossby and Flagg, 2012}.

2.2. Satellite Altimetry

In this study, we use the Mercator gridded (1/3° × 1/3°) daily sea surface heights (SSHs) in form of absolute dynamic topography (= sea-level anomalies + mean dynamic topography). This product distributed by AVISO (Archiving, Validation and Interpretation of Satellite Oceanographic data, http://www.aviso.oceanobs.com) through the SSALTO/DUACS project. Most recently, \textit{Rio et al.} \cite{2011} showed an improved marine geoid and thus the SSH now provides a better consistency with independent in situ measurements. It is worth mentioning that the mean dynamic topography \textit{Rio et al.} \cite{2011} is based on several data sets: geoid model, an altimetric Mean Sea Surface as well as altimetric Sea-Level Anomalies, wind stress data and oceanographic in situ data. In order to ensure that the sampling is stable in time during the analysis period (1993–2010), we use the “Reference” product of the SSH, i.e., two satellites orbiting over the same path over the entire sampling period.

The daily SSH has been monthly averaged and thereafter deseasoned in order to remove the strong seasonal cycle in the sea-surface height. The analysis focuses on a 1° wide swath section between 59° and 60° N from the Irminger Sea, across the RR and the Iceland Basin to the Scotland Slope. The grid points between 59° and 60° N have not been averaged in the north-south, but along a line parallel to the RR.

3. Results

3.1. Organization of Flow

We begin by considering northward flow from just east of the East Greenland Current to Hatton Bank (cf. Figure 1), the westemmost of the Banks west of Scotland. We adopt the same approach as K-etal, who noted that the RR very effectively organizes mean flow parallel to the ridge. They showed this in a comparison of the velocity structure of the constant latitude and great circle routes (from now on referred to as the

\begin{figure}
\centering
\includegraphics[width=\textwidth]{figure1.png}
\caption{Bathymetric chart of the northeast Atlantic Ocean with Greenland in the west, Iceland in the center, and Scotland in the east. The thin gray lines indicate the available Nuka Arctica ADCP data. West and eastbound transits take place along great circles and constant latitudes, respectively. The large dashed box delimits the data used for the central basin analysis to highlight the dynamical constraint of the Reykjanes Ridge. The 2° inner box together with the narrow wedge in the east delimits the data used for the full zonal analysis. The short red lines indicate 500 km steps from the ridge crest. The acronyms represent (from east to west) the Maury Channel (MC), the Hatton Bank (HB), the RB (Rockall Bank), and the George Bligh Bank (GBB).}
\end{figure}
C and G-routes) relative to the ridge axis (long red line in Figure 1). Given this, we collapse all data from both groups into a single zonal-vertical section organized by zonal distance from RR in Figure 1. This zonal organization is further aided by the southwest-northeast orientation of the East Greenland Current, the Maury Channel in the Iceland Basin and Hatton Bank (all no doubt due to seafloor spreading). From hereon zonal distance will be in kilometer from the RR ridge crest, the long red line in Figure 1. In section 3.3, we extend the analysis to the full section, but due to the complexity of flow over the Banks we restrict that analysis to the C-group of sections only.

We begin by showing a typical zonal ADCP section in Figure 2. This represents the “raw” information that will be used in the following zonal averages. We simply show the observed velocity at three depths, 23 m, which includes the ageostrophic Ekman layer, 151 m representing the upper ocean and 351 m at about the maximum depth the 150 kHz ADCP could reach. Notice how the velocity vectors can point in any direction, reflecting the energetic mesoscale eddy field. There is a hint that the 151 and 351 m vectors are more closely aligned than either is with the vector at 23 m, which is likely influenced by the overlying wind field. This single section serves as an effective reminder that it tells us little about the underlying mean velocity field, as will become evident below.

Following K-etal, we show topographic and basin constraint on the velocity field by rotating the mean velocity field into components parallel and orthogonal to the RR using all data inside the large dashed box. The C sections (somewhat more in the far west) available for this define the number of degrees of freedom (DOF) since the section the Nuka Arctica operates on a 21 day schedule. Figure 3 shows the suppression of the normal component of motion (top) by steering the mean flow along the topography (middle). The figure reveals the fine structure previously noted by K-etal: the two flows to the north on the western slope and one to the south on the eastern slope. The core of this southward flow is not just east of the ridge crest, but some 100 km farther east roughly over the 2 km isobath where the south flowing Faroe Bank overflow water is located (Borenás and Lundberg, 1988). There is also a well-defined flow north at +500 km corresponding to the Maury Channel in the Iceland Basin, and at approximately −300 km in the Irminger Sea. The overall ratio of mean kinetic energy parallel and normal to the RR (in the top 400 m between 600 and 700 km) exceeds a factor 7. The southward flow on the eastern side is even more tightly constrained to follow bathymetry such that the ratio of parallel to normal mean kinetic energy increases to about ~50:1. The bottom plot shows the mean bathymetric profile for this domain averaged along lines parallel to the RR.

3.2. Poleward Transport Between Scotland and Greenland

The previous section used all sections in the large dashed box in Figure 1; we now use only the zonal sections inside the smaller zonal box and the narrow wedge in Figure 1. Although this reduces the DOF roughly in half to ~20, this restriction removes transits over Banks both to the north and the south and over Hatton Bank. The western extent of the figure is the same as in Figure 2, but extends here to the Scotland shelf in the east. The intent is to create a more homogenous data set with respect to the complex bank topography; thus the wedge stretches east just north of Hatton Bank between Lousy Bank to its north and George Bligh Bank and Rosemary Bank to its south. The mean velocity field of this section, now in geographic coordinates is shown in Figure 4. Perhaps the most striking feature is the strong zonal flow between 650 and 850 km (top) due to the eastward extension of Hatton Bank. The meridional flow has two (less distinct)
maxima as was seen in Figure 3 west of the RR, at the Maury Channel at +450 and +600 km, at +900 km just beyond the eastern end of George Bligh Bank, and along the Scotland slope, also known as the Slope Current. The bottom plot shows the bathymetric profile between 59° and 60° N averaged parallel to the RR. Figure 5 shows ADCP-measured eddy kinetic energy (EKE) along the section (top). The EKE is high at +500 km, the Maury Channel. Not only is it only much higher than anywhere else, it also appears to be remarkably localized [Martin et al., 1998]. Unlike the mean kinetic energy which is strongly shaped by bathymetry the EKE field is nearly isotropic, only slightly more energetic (~10%) over and parallel to the RR bathymetry.

Figure 3. Mean velocity (ms⁻¹) (top) normal and (middle) parallel to Reykjanes Ridge as a function of distance from the ridge axis. (bottom) The ridge-crest parallel mean topography.

Figure 4. Mean (top) zonal and (middle) meridional velocity (ms⁻¹). (bottom) Average bathymetry between 59° and 60° N averaged parallel to the Reykjanes Ridge.
EKE appears to decay more rapidly with depth west of the RR than to its east, probably reflecting the shallower pycnocline there. The EKE is slightly higher over the western slope than the eastern slope of the RR, but the difference is not as compelling as the altimetric EKE (middle and bottom), which in fact reveal a minimum along the ridge axis (dark blue color) as well as between Hatton Bank and Lousy Bank, perhaps due to flow stabilization by the trough in between. The overall mean of EKE just below the wind-energized surface waters is 0.023 m$^2$ s$^{-2}$. While higher, this compares reasonably well with the overall geostrophic EKE we estimate for the same zonal band from the spatially smoothed AVISO product.

### 3.3. Transport

To estimate poleward transport through this zonal section, we use two definitions, both based on the meridional or south-to-north component of motion. The first is the zonal integral of meridional velocity at the surface (1 m thick layer at 23 m depth). We call this layer transport. The reason for this integral is that through geostrophic scaling it can be used to determine the required dynamic height or equivalently pressure difference expressed in units of meters of water $\Delta H$ to balance that flow:

$$\Delta H = \frac{f}{g} \int v dx$$  \hspace{1cm} (1)

where $f$ and $g$ represent the Coriolis parameter and acceleration due to gravity, respectively. $\Delta H$ can thus be compared with variations in sea surface height ($\Delta SSH$) estimated from satellite altimetry. Of course, the same integral can be used to estimate dynamic height difference at any depth. The second definition of transport takes advantage of the ADCP’s ability to profile currents to $\sim$400 m, the maximum useful
operating depth of the 150 kHz instrument. This gives us a direct estimate of volume flux between Greenland and Scotland to this depth.

Integrating velocity over short distances, O(10^2) km, to determine transport can be done with some confidence, but when integrating velocity over greater distances any slight bias in velocity due to heading error will eventually render the integration useless. In Appendix A, we discuss the question of measurement and integration error in some detail. We conclude there that the overall uncertainty in transport from the ensemble of sections is \( \frac{\text{C24}}{\text{C24}} \). While this is not accurate enough to quantify interannual variations in basin-wide transport, the data reveal clearly the principal transport paths, their partition west and east of the RR, and between the Iceland Basin and the Banks area. Interannual variations will be examined in the next section.

Figure 6 shows the 0–400 m vertically averaged velocity and transport relative to the RR west to the East Greenland Current and east to the Scotland shelf. The poleward transport is almost the same on both sides of the ridge, 8 ± 1.6 Sv east and 8.5 ± 1.5 Sv west of the RR. Of the western branch, 5 Sv takes place over the RR slope within 150 km of the ridge crest with the remainder more uniformly distributed across the Irminger Sea. To the east of the ridge, the flow is concentrated to three steps: 450 and 600 km (2.5 and 2 Sv), 900 km (1.2 Sv), and 1300 km (1.7 Sv), all associated with topography. The first is located over Maury Channel, the deepest part of the Iceland Basin, the second just east of George Bligh Bank (which not only cuts into the southern half of the wedge at 14°W, but is semiconnected to the longer Rockall Bank to its south) and the third is on the Scottish slope. The transport curve suggests that little transport occurs elsewhere.

3.4. The Altimetric Perspective

We begin by asking how well \( \Delta \text{SSH} \) from altimetry and geostrophic sea-level \( \Delta \text{H} \) from the ADCP agree. For this purpose, we construct a time series of \( \Delta \text{SSH} \) sections of between 59° and 60°N from the East Greenland Current to the Scotland shelf (but no wedge taper in the east). As before, we will use the RR axis as the anchor point for the comparisons, justified by the fact that on average the along ridge flow changes sign there. The resulting curves are shown in Figure 7. The two \( \Delta \text{SSH} \) curves represent the full 18 year record (orange) and the 1999–2002 period (black), respectively. The tightness of agreement between the two
Figure 7. Dynamic height relative to RR estimated from the ADCP sections along the 59.5°N at two depths (20 and 55 m) averaged over 1999–2002 period, as well as the difference in the SSH (absolute dynamic topography) for both the averaged 1999–2002 and the 1993–2010 altimetric record.

Figure 8. Hovmöller diagram of deseasoned SSH (absolute dynamic topography) as averaged parallel to the RR between 59° and 60°N from across the Irminger Sea in the west to the Scotland slope in the east. The contour spacing is 0.1 m.
suggests that the shorter period is typical, and reflects the full monitoring period rather closely. The overall agreement with $\Delta H$ is quite good. Across the Irminger Sea, the agreement is within 0.05 m everywhere. However, east over the Banks area between 1000 and 1200 km $\Delta SSSH$ exceeds $\Delta H$ by 0.15 m and is still 0.1 m greater at the Scotland shelf. Interestingly, the ADCP reveals a negative shear in the top ten's of meters in this region (~600–1200 km, Figure 3), perhaps due to a strong Ekman flow driven by persistent westerlies west of Scotland. Since Ekman flow has no pressure signal associated with it, it should not be included in the sea-level integration. An ad hoc avoidance of the Ekman layer by integrating velocity just below the sheared surface waters, at 55 m say, gives us a $\Delta H$ (the blue curve) in much better agreement with $\Delta SSSH$.

(We return to the question of Ekman layer flow in the discussion.) At the mesoscale level, the agreement between $\Delta SSSH$ and $\Delta H$ is less satisfactory although major transitions in $\Delta H$ can be associated with the corresponding $\Delta SSSH$ signals. These include the northward flow just west of the RR (100 km), over the Maury Channel (450 km), at George Bligh Bank (950 km), and the Scottish slope (1300 km). The southward flow east of the RR (100 km) does not show up in $\Delta SSSH$ although a leveling out is evident. The limitation in spatial resolution notwithstanding, the overall agreement between $\Delta H$ and $\Delta SSSH$ is encouraging, and one can expect further improvement as the accuracy of the geoid (used in the calculation of the absolute dynamic topography) improves.

3.5. Temporal Variability
We now turn to the 18 year long SSH time series to examine how poleward surface fluxes vary with time. Figure 8 shows the difference in the SSH in a 1° latitude zonal strip centered at 59.5° N. Since, there are AVISO data points every 1/3° latitude we average these along a line parallel to the RR as we did with the ADCP data. This maintains close correspondence between the ADCP and $\Delta SSSH$ data.

One can see a general increase in the SSH over the 18 year record, both in the east and the west such the zonally averaged increase in $\Delta SSSH$ over the full period is 0.08 m, which is equivalent to 4 mm/year. But here we are interested in temporal changes in poleward transport, i.e. changes in zonal sea-level difference. As Figure 7 indicates, there are longitudes where $\Delta SSSH$ exhibits rapid increases, such as at about −100, +500, and +1000 km. The first is near the RR, the second over the Maury Channel, and the third near George Bligh.

Figure 9. Same as Figure 8 except that each zonal section has been zeroed at the Reykjanes Ridge. The contour spacing is 0.1 m.
Seamount. The ADCP sections tell us that the RR acts not only as a separator of flow to the Nordic and Labrador Seas, it serves also as a “zero” streamline in the sense that the mean flow is in opposite directions to either side. This is of course not true in the instantaneous sense given ever-present eddy activity, but it applies on a yearly basis as can be seen from the four 1 year ADCP averages (Figure A1). While not proven, we assume this applies to yearly averages in general, because, as we shall see, it is convenient to continue to use the RR as a reference point to highlight along section variability more clearly, Figure 9. Note that this RR-referencing has no impact on estimation of meridional flow, which depends only upon zonal SSH differences. (It does remove the gradual increase in SSH over time.) Interestingly, the southward flow east of the ridge is more pronounced in 1996, 1997, and 2010, perhaps in response to the very negative NAO winters at the time.

This reorganization reveals more clearly areas where SSH exhibits preferential steps such as near 500 and 1000 km; it also hints at a reverse slope east of the RR. The RMS variability of SSH across the ocean is 0.03–0.035 m except locally at 500 km (20°W) where it nearly doubles to 0.07 m (this coincides with the highly elevated EKE in Figure 5). While SSH still has difficulties resolving the mean field on smaller scales, its ability to resolve long-term change is excellent even if the accuracy of the sea-level anomalies is around 2–3 cm [Robinson, 2004]. The question we explore here is long-term change along this section. The RR serves as separator of flow to two quite different areas of winter-time cooling and convective overturning: the Labrador Sea with its production of North Atlantic water of intermediate density, and the Nordic Seas with its production of dense overflow water. An obvious question is how do these (surface) transports (expressed as a SSH difference) vary, and do they give any indication of long-term trend? Just as with transport (Figure 6) surface fluxes west and east of the RR as measured by SSH differences have about the same overall amplitude, 0.3 m, but SSH reveals that they vary almost perfectly out of phase such that when the Irminger Sea flux increases, the Nordic Seas branch decreases such that the sum of the two, Scotland Slope-Irminger Sea, shows no evidence of this variability (Figure 10). If we assume that the 0–400 m transport...
scales the same way, the $\sim 0.02$ m amplitude would correspond to a $0.02/0.3*8 \sim 0.5$ Sv variation in transport. The purpose of this calculation is to provide a measure of what might be variations in transport. What is actually taking place will require further study. The Nordic Seas branch would have a similar transport variation, but with opposite phase. Similarly, the Nordic Seas branch can be divided into two subbranches, one via the open Iceland Basin ($\Delta SSH \sim 0.17$ m), and one through the Banks region ($\Delta SSH \sim 0.12$ m) such that a $\sim 0.03$ m amplitude would correspond to roughly 0.9 and 0.8 Sv, respectively. Interestingly, all these variations cancel out leaving a rather well-defined decrease in overall sea-level difference of 0.06 m over the 18 year observing period. This decrease has comparable contributions from both branches. This drop, applied to the top 400 m transport would correspond to a 1.7 Sv decrease in poleward transport.

4. Discussion

4.1. Spatial Patterns

The combination of ADCP scans of ocean currents and satellite altimetry both point to the pivotal role played by the RR: the former reveals a striking organization of currents, the latter in terms of how the RR partitions temporal variability. The high-resolution sampling by the Nuka Arctica in both the horizontal and the vertical measures the mean velocity field with a resolution unavailable by any other technique. The feature that shows up most prominently in these sections along 59.5°N must be the flows parallel to the ridge: the two flows to the north and the flow south, west and east of the RR crest, respectively; both noted earlier by K-etal. The 5 Sv transport flow north along the western slope suggests that this is a major contributor toward the Labrador Sea. The weak but well-defined flow south just east of the crest appears to be the surface manifestation of the strong flow south at depth from the Faroe Bank Channel overflow. The repeat sampling also reveals two distinct flows north over the deep Maury Channel (at 500 and 600 km) separated by a narrow band of southward flow. This pattern is also accompanied by highly localized energetic EKE consistent with other observations of intense eddy activity over the deepest part of the basin. Thus, the Maury Channel both funnels a significant amount of water north as well as acts as an attractor for eddy activity. Martin et al. [1998] suggest that the coherent eddies they have observed here may have spun off the warm Subpolar Front as it curves around the southern end of Hatton Bank. A localized concentration of high EKE over the deepest part of a basin has also been observed in the Lofoten Basin; in that case due to
the migration of anticyclonic eddies into the deepest part of the basin [Köhl, 2007; Rossby et al., 2009; Koszalka et al., 2011].

The southward flow just east of the RR crest has largely gone unnoticed in the earlier literature, but the Jakobsen et al. [2003] analysis of surface drifter data shows it clearly (see also Reverdin et al. [2003]). This <2 Sv southward transport very likely results from the strong cyclonic circulation at depth in the Iceland Basin [Bower et al., 2002]. A possible reason it had not been recognized may be that the density field shoals from east to west across the RR giving rise to a positive baroclinic shear, which would bespeak a northward flow at the surface. But unless one knows the strength of this southward flow at depth any geostrophic estimate of flow at the surface will be too positive. Very likely most of this southward flow crosses the RR farther south and turns north, but it is too weak to be the main source of water that has been postulated to cross the RR and flow north along the western slope [Brambilla and Talley, 2008]. Instead most of that water must come from the Subpolar Front crossing the mid-Atlantic ridge from west to east in the vicinity of the Charlie Gibbs/Faraday Fracture Zones and then retroflecting back toward the RR [Bower et al., 2002]. Willis and Fu [2008] in their altimeter/Argo float construction of the North Atlantic circulation also show a significant retroflection of the North Atlantic Current south of 60°N. Indeed, it seems plausible that the two northward flows originate at specific fracture zones such as at 57° and 55°N [cf. Xu et al., 2010; A. Bower, personal communication 2013]. Well to the east of the RR poleward flow is clearly concentrated to several sites that are probably topographically defined although the local dynamics may differ. Thus two mean flows, at 450 and 600 km, lie right over the deepest part of the Iceland Basin, the Maury Channel, and not, as one might expect on the basis of the flow north along the RR, pressing against the western slope of Hatton Bank. Farther east at 900 and 1300 km there are two additional contributions to northward transport, similar in magnitude: the former appears to be coming north along the eastern side of George Bligh Bank, part of the Rockall Plateau complex, and the latter along the Scotland slope. With the exception of the Slope Current, these localized mean flows may not show

Figure 12. Poleward surface flux relative to mean velocity in the 60–90 m depth range (thin gray line). The heavy line indicates its low-pass filter using a fourth-order Butterworth filter with a 50 km cutoff.
up as such in any single transect. It is through repeat sampling that this underlying mean field can be mapped and quantified.

In section 3.3, we reported on the distribution of northward transport in the top 400 m. The uncertainty in the northward transport between the RR and the Scotland Slope based on about 20 contributing sections is estimated to be 1.6 Sv although there is an additional heading uncertainty bias we cannot determine (Appendix A). Relative to the total transport, 8 Sv, this is a significant uncertainty. On the other hand, it turns out that this transport is of the same magnitude as the $8.5 \pm 0.3$ Sv direct measurement of transport between Iceland and Scotland reported by Rossby and Flagg [2012]. A more detailed comparison would require examination of flow in those density classes entering the Nordic Seas, but it does suggest that the George Bligh and Slope Currents continue preferentially toward the Nordic Seas. We also think that most of the near-surface water in Maury Channel branch continues toward and continues across the Iceland-Faroe Ridge with remainder curving west and south along the eastern slope of the RR. How much is sent in each direction must depend heavily upon density and sea-level differences across the Iceland-Faroe Ridge in the north and the RR to the west. The absence of a significant transport south along the eastern slope of the RR would be consistent with this. Further evidence for a lack of cross-RR flow north of the Nuka Arctica line can be seen in Figure 11, which shows the deseasoned and low-pass filtered (using a 1 year running mean) Hovmöller diagram of SSH along the RR crest. SSH shows almost no gradient between $60^\circ$ and $62.5^\circ$N and varies in-phase along the entire section. To the south of $60^\circ$N one sees a clear decrease in SSH. This would accord with the idea that the northward flow on the western side of the RR is fed by water retroflected from the east flowing North Atlantic Current farther south as mentioned earlier. Just as important, the lack of any SSH gradient along the RR north of $60^\circ$N is further evidence of the strong topography constraint exerted by the RR such that any cross-ridge transport there must be minimal and limited principally to the Ekman layer.

The $8.5$ Sv transport north west of the RR comprises both about $5$ Sv flowing along the western slope of the RR and $3.5$ Sv more or less uniformly distributed across the Irminger Sea up to the offshore edge of the East Greenland Current [Våge et al., 2011]. This program cannot by itself tell whether any of this $3.5$ Sv interior flow is recirculated within the gyre, but it would seem likely based on the conspicuous doming of the density field there [Sarafanov et al., 2012]. A fraction of the flow north along the RR probably feeds the West Iceland Current [Jónsson and Valdimarsson, 2005]. The alternative would be a diversion of some Maury Channel flow west across the ridge north of the Nuka Arctica line, but given the strong constraint imposed by the RR as noted in the previous paragraph this seems rather unlikely.

We wrote in section 3.4 that we first thought that the flow south east of +600 km be an Ekman layer flow, because it is so clearly concentrated to the surface. However, this would require that the winds are similarly well organized in that area compared to the rest of the ocean between Greenland and Scotland, but we find no evidence in support this assumption [Isemmer and Hasse, 1987]. It is also curious that this surface flow occurs principally over the Banks region. To estimate the magnitude of this surface flow, we postulate that

![Figure 13](image-url). A schematic representation of 0–400 m poleward transport across the northeast Atlantic between Greenland and Scotland. The flow north along the western slope of the RR is primarily destined for the Labrador Sea, and while transport toward the Nordic Seas appears to be distributed between the Iceland Basin west of Hatton Bank and the Scotland slope with a smaller amount flowing north along the eastern margin of Rockall/George Bligh Banks. A weak recirculation occurs in both the Iceland Basin and Irminger Seas; to what extent they are closed is unclear.
the flow below 60 m is essentially geostrophic. Thus, we first approximate the underlying geostrophic flow as the mean flow between 60 and 90 m depth. This is subtracted from the top 60 m to give us the excess surface transport. This is plotted in Figure 12 as the thin curve with data points every 5 km. A low-pass filter with 50 km cutoff wavelength results in the heavy curve. Note how everywhere west of about ±500 km the mean transport fluctuates around 0. But east of 500 km and especially between 800 and 1150 km, we find an exceptionally strong flow to the south followed by a narrow flow to the north just east of 1200 km. The net surface transport as estimated here about 0.34 Sv to the south. We do not know the reasons for these concentrated surface flows just over the Banks region.

4.2. Temporal Variability

Both the regularity and complementarity of altimetric variability west and east of the RR, as well as west and east of Hatton Bank come as a complete surprise. That they are nearly perfectly out of phase means that the total transport north remains relatively stable might suggest that these variations might be internal basin modes coupled along the RR ridge. The fact that the signals are similar in size to either side of the RR implies comparable transport variations. Since, to our knowledge, there is no evidence that this signal continues into the Nordic Seas, it would imply that the two basins buffer and trap the signal there. We hypothesize that the NAC splits or bifurcates into two branches, one toward the Iceland Basin and the other toward Irminger Sea, with the southern end of the Reykjanes Ridge acting as the "splitter." As water flows through the basins, there is an imbalance such that some backs up in the Irminger Sea rather than continues toward the Labrador Sea (via the boundary currents around Greenland). This "backing filling" (superimposed on the through flow) causes the flow from the NAC to weaken toward the Irminger Sea with a slight weakening of sea-level difference across it as a consequence, and increase toward the Iceland Basin instead until it "back fills" at which point transport increases toward the Irminger Sea. This effect is readily achieved by letting sea level along the RR to gradually drop and rise. Curiously, Figure 10 suggests that the Iceland Basin exhibits a similar oscillatory behavior. Perhaps the Iceland Basin branch undergoes a similar bifurcation at the southern end of Hatton Bank leading to a similar “oscillation” west and east of it. These “oscillations” do not appear to be (tightly) coupled to the NAO.

The long-term decrease in sea-level tilt has been noted before [see e.g., Häkkinen and Rhines, 2004; Zhang, 2008]. Whereas they note a gradual rise in SSH across the subpolar gyre, we also note a weakening sea-level tilt all along 59.5°N, from off Greenland to Scotland, i.e., across both the Irminger Sea and the Iceland Basin. A linear fit to this trend leads to an overall decrease of 0.06 m, which relative to the total sea-level difference of ~0.6 m implies a 10% decrease. Applying this to the 16.5 Sv for the 0–400 m average transport would mean a ~1.7 Sv reduction in transport, roughly equally partitioned between the Nordic and Labrador Seas. Evidently, the observed trend noted in those studies continues unabated through the end of 2010, an 18 year sustained decrease, a time scale significantly longer than that of the NAO, and perhaps associated with the Atlantic multidecadal oscillation [Zhang, 2008]. In this context, it is worth recalling that the Nuka Arctica program was restarted in fall 2012, and as the number of sections increase, we may be in a position to examine or test for possible long-term changes in temporal and zonal distribution of poleward transport.

5. Summary

The reprocessed 1999–2002 set of ADCP sections between Greenland and Scotland have enabled a detailed analysis of transport across a zonal line centered at 59.5°N. The velocity field in the top 400 m reveals that the Reykjanes Ridge serves as a very effective separator of flow toward the Nordic and Labrador Seas, respectively. Clearly, topography plays a fundamental role in guiding these two branches, but whereas the Labrador Sea branch exhibits two well-defined flows to the north, both on the western slope of the Reykjanes Ridge, the eastern branch flows north in roughly equal amounts, one over the deep Maury channel west of Hatton Bank, and the other with roughly equal contributions from just east George Bligh Bank and the Slope Current, respectively, Figure 13. There is also a weak but rather persistent flow south along the eastern slope of the RR. We have sketched this as a local recirculation, almost certainly a surface manifestation of the strong cyclonic circulation at depth [Bower et al., 2002] driven by the Faroe Bank overflow. Whether this surface circulation actually has a sea level minimum in the Eulerian sense is unclear, hence the dashed line (Figure 13). Similarly, the general circulation in the Irminger Sea is cyclonic, but here too it is not
clear whether it has a closed recirculating core. We also conclude that there is very little mean flow across the RR from the Nuka Arctica line north toward Iceland, both because of the evident constraint imposed by the ridge and because of the lack of any SSH gradients along the ridge to the north. To the south of the Nuka Arctica line, one does find decreasing along-ridge sea level suggesting significant east-west cross-ridge flow there. This would be consistent with the idea that the northward flow on the western side of the RR is fed by the east flowing NAC retroreflecting and crossing the RR to the west well to the south of 60°N (not shown). The similar transport values between those obtained here and at the Iceland-Faroe-Scotland ridge as measured by the Norrøna support the idea that the Maury Channel flow feeds the Iceland-Faroe ridge, whereas the two eastern subbranches feed the flow through the Faroe-Shetland Channel (Chafik, 2012). The numbers should be viewed as approximate since these transports reported here have not been remapped into the density coordinates used in the Norrøna study (Rossby and Flagg, 2012). While there are significant uncertainties in integrating transport over such large distances, to the extent that much if not most of the flow is concentrated to narrow topographically confined regions, the ADCP estimates of transport in these confined regions will be considerably more accurate since the instrumental error over short distances is limited [Worst et al., 2013]. These are questions that will be the subject of future studies.

The satellite altimetric absolute dynamic topography data show good overall agreement with geostrophically determined sea-level difference from the repeat ADCP sections, but are unable to fully resolve the fine structure of the topographically defined mean circulation. On the other hand, by partitioning the 59.5°N section into two subsections defined by the RR, the excellent altimetric coverage reveals that poleward flow west and east of the RR is strongly anticorrelated. The data also reveal that the two eastern subbranches, west and east of Hatton Bank, are also strongly anticorrelated, but offset in time with respect to the Labrador Sea branch. Remarkably, all these variations largely cancel out for the entire Greenland-Scotland section leaving a gradual decrease in sea-level difference from 0.62 to 0.56 m over the 1993–2011 observation period. This, if real, would point to a 10% overall weakening of the MOC in the northern North Atlantic, roughly equally divided between the Labrador and Nordic Sea branches.

The Nuka Arctica ADCP program resumed operation in fall 2012, now with a 75 kHz instrument that reaches to about 800 m. As this new database grows it will become possible to examine interdecadal change in the upper ocean of the northeast Atlantic. This was a major motivation for the reprocessing of the earlier data—to ensure that they have been fully examined and properly archived.

Appendix A: Ocean Current and Transport Error Discussion

A1. Instrumental Errors

To determine geo-referenced water velocity from a moving platform requires two measurements, water velocity relative to the ship (the ADCP) and ship motion over the bottom (vessel speed and heading, both from GPS).

According to the Teledyne RDInstruments, the manufacturer of the 150 kHz ADCP used here, the standard deviation of the 5 min ensemble average (16 m pulse length, 1 ping per second = 300) is 0.0038 ms\(^{-1}\) (RDI SC-ADCP technical manual page 2–11, September, 1992). This is the uncertainty of the 5 min ensemble estimate of velocity measured by the ADCP. It is small because of the high operating frequency (150 kHz) and the large number of samples (300) in each 5 min ensemble.

GPS-compass (Thales ADU-5): The instantaneous heading error for a 5 m antenna separation is listed to be about 0.05° adding a 0.007 ms\(^{-1}\) random error to the instantaneous velocity field. This uncertainty has a very fast decorrelation time and according to the manufacturer is not influenced by vessel motion, although there may be times when certain satellite configurations could degrade the accuracy briefly. Given a full decorrelation time due to vessel motion is \(\sim 10\) s, the 5 min standard error in velocity due to heading uncertainties would therefore be the instantaneous velocity uncertainty divided by number of degrees of freedom \(\sqrt{300/50/\text{period}}=0.0013\) ms\(^{-1}\). While a bit smaller than the uncertainty of the ADCP measurement it adds to a combined ADCP and GPS heading uncertainty of about 0.004 ms\(^{-1}\); this is the resulting measurement uncertainty on a moving vessel. The more serious challenge is to determine the physical relationship between the ADCP and the GPS compass so that vessel speed can be removed from the acoustic measurement to obtain unbiased geo-referenced velocity.

Conceptually, the approach is straightforward: we operate the ADCP in bottom-track mode to integrate measured bottom “speed” to determine distance and direction traveled and compare this with GPS...
estimate of distance traveled. This comparison gives us the required information to align the ADCP correctly as well as check on its gain factor. The overall standard deviation of these calibrations is about 0.07 or equivalently about 0.01 ms\(^{-1}\). This is the transit-to-transit uncertainty in cross-track velocity uncertainty. It is our biggest concern. For studies of eddy variability in the ocean this is an acceptably small number, for determining transport over large distances it is not.

In this study, we apply the calibration obtained while the vessel is transiting the shallow North Sea to the remainder of that transit to/from Greenland. We could alternatively take the ensemble average of all calibrations and apply this to all transits. On the assumption that the physical relationship between the GPS-compass and the ADCP remains absolutely invariant this would be best. But for reasons we cannot explain, we have in the postprocessing discovered a drift in the bottom track calibration (years 2001 and 2002) that cannot be dismissed. This means that each transit may have a cross-track velocity bias of about 0.01 ms\(^{-1}\). This is a serious limitation. However, assuming these calibration uncertainties are normally distributed, the residual bias from the ensemble of 20 sections (for the long section between Greenland and the Scotland slope) used here would \(0.01/\sqrt{20} = 0.0022\) ms\(^{-1}\). This bias, assuming it applies equally to the full length of the 1900 km long section would lead to transport uncertainty of \(400 \text{ m} \times 1900 \text{ km} \times 0.0022 \text{ ms}^{-1} = 1.67\) Sv. This would be the residual irreducible uncertainty of transport integration for the full Nuka Arctica section. The corresponding uncertainty in layer transport would be \(0.0022 \text{ ms}^{-1} \times 1900 \text{ km} = 4180 \text{ m}^2 \text{s}^{-1}\) or in geostrophic terms = 0.054 m sea-level difference over the 1900 km distance. For shorter sections, these uncertainties will be reduced accordingly. This demonstrates the criticality of knowing the orientation of the ADCP accurately.

A2. Oceanic Uncertainties

Characteristic eddy speeds (\(v'\)) along the Nuka Arctica section vary range between 0.1 and 0.15 ms\(^{-1}\) except in the center of the Iceland Basin where they increase to \(\sim 0.22\) ms\(^{-1}\) (Figure 5). This eddy activity introduces increases the “noise level” when estimating transport. The more “noise” we can average over the better

![Figure A1](image-url)
we can reduce its effects. The scale of the meso-scale eddy field is small and can be characterized by the half-power point of the transverse velocity component (~12 km along the Nuka Line) alternatively the radius of deformation, which provides a good measure of eddy scale (~10 km; [Chelton et al., 1998]). This distance divided into the integration length gives a measure of the degrees of freedom (DOF) available. Thus for the 1900 km section, we have ~1900/12 = 158 DOF when integrating the full Nuka Arctica section. Taking 0.14 ms⁻¹ as an average v' the corresponding standard error for any integration would be about 0.14/√158 = 0.011ms⁻¹. This velocity uncertainty times the cross-sectional area (length of the section × depth) gives the single section uncertainty in transport (in this case ~8.5 Sv for the upper 400 m transport). It is the repeat sampling that adds the crucial DOF to reduce this uncertainty. Ensemble averaging over the N = 20 sections results in an oceanic uncertainty of 1.9 Sv or 12% of the mean. To summarize: the transport uncertainty = v × length of section × depth/DOF × N.

Applying these numbers to the sections for each of the 4 years (N = 10) in the larger box (~600 to +700 km) the uncertainty in annual transport for the Irminger Sea and Iceland Basin is 1.5 and 1.6 Sv, respectively. For the 4 year average, these numbers will be half as large (~40 versus 10 sections). These numbers are in reasonable agreement with observation, Figure A1.

References


