North Atlantic meridional overturning circulation variations from GRACE ocean bottom pressure anomalies

Felix W. Landerer1, David N. Wiese1, Katrin Bentel1, Carmen Boening1, and Michael M. Watkins1,2

1Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, USA, 2University of Texas at Austin / CSR, Austin, Texas, USA

Abstract

Concerns about North Atlantic Meridional Overturning Circulation (NAMOC) changes imply the need for a continuous, large-scale observation capability to detect changes on interannual to decadal time scales. Here we present the first measurements of Lower North Atlantic Deep Water (LNADW) transport changes using only time-variable gravity observations from Gravity Recovery and Climate Experiment (GRACE) satellites from 2003 until now. Improved monthly gravity field retrievals allow the detection of North Atlantic interannual bottom pressure anomalies and LNADW transport estimates that are in good agreement with those from the Rapid Climate Change-Meridional Overturning Circulation and Heatflux Array (RAPID/MOCHA). Concurrent with the observed AMOC transport anomalies from late 2009 through early 2010, GRACE measured ocean bottom pressures changes in the 3000–5000 m deep western North Atlantic on the order of 20 mm H2O (200 Pa), implying a southward volume transport anomaly in that layer of approximately –5.5 Sverdrup. Our results highlight the efficacy of space gravimetry for observing AMOC variations to evaluate latitudinal coherency and long-term variability.

1. Introduction

The Atlantic Meridional Overturning Circulation (AMOC) plays a key role in the poleward transport of heat. Changes in this transport can influence climate at higher latitudes significantly, with potentially significant impacts for the Northern Hemisphere, in particular northwest Europe’s climate [Manabe and Stouffer, 1995; Srokosz et al., 2012; Intergovernmental Panel on Climate Change (IPCC), 2013]. The dynamics of the mean and time-variable North Atlantic MOC across latitudes have been described in several model studies [e.g., Vellinga and Wood (2002); Bingham and Hughes (2008, 2009), hereafter BH08 and BH09; Roussenov et al., 2008; McCarthy et al. (2012); Wunsch and Heimbach (2013)]. Continuous in situ observations with hydrographic arrays such as Rapid Climate Change-Meridional Overturning Circulation and Heatflux Array (RAPID/MOCHA) [e.g., Kanzow et al., 2007; Elipot et al., 2014] and Meridional Overturning Variability Experiment (MOVE) [Send et al., 2011] exist for about a decade now but are generally limited to one particular latitude. Another AMOC estimate from Willis (2010) uses in situ hydrographic profiles in combination with satellite sea surface height observations to derive the strength of the northward upper AMOC cell at 41°N, and Frajka-Williams (2015) recently presented an extended data record back to 1993 at 26.5°N by statistically combining hydrographic and satellite sea surface height data. These studies have also led to a more refined understanding of AMOC variations across several space and time scales. While the aforementioned observational estimates all detected a prominent AMOC slowdown of more than 5 sverdrup (Sv) in late 2009 and early 2010 [e.g., Srokosz et al., 2012], the notion of a single, latitudinally coherent cell has been challenged in both observations [e.g., Lozier et al., 2010] and model studies [Bingham et al., 2007; Wunsch and Heimbach, 2013]. Instead, AMOC changes can be latitude dependent and gyre specific [Lozier et al., 2010] and can have different character north and south of approximately 40°N [Bingham et al., 2007]. The evolving understanding of AMOC variability, in combination with concerns about potential future AMOC decreases [e.g., IPCC, 2013; Rahmstorf et al., 2015], highlights the need for observation methods that can observe AMOC changes continuously in both time and space.

Here we describe AMOC variability detected in ocean bottom pressure (OBP) anomalies that are inferred from time-variable gravity observations from the Gravity Recovery and Climate Experiment (GRACE) satellites [Tapley et al., 2004]. The GRACE gravity observations have provided complete global spatial coverage of ocean mass changes on monthly time scales from 2003 until present [e.g., Chambers and Bonin, 2012]. While the feasibility of using bottom pressure to derive AMOC variations has been demonstrated in
2. Methods and Data

The AMOC consists of a northward flow in the upper layer of the ocean (mostly between the surface and 1000 m depth, e.g., Wunsch and Heimbach [2013]), and a return flow of North Atlantic Deep Water (NADW) to the south in the deeper layers between 1000 and 5000 m. Large-scale flows are generally in geostrophic balance, and the meridional transport $T(y, z)$ at a particular latitude $y$ and depth $z$ can be derived from the zonal bottom pressure differences $p_E$ and $p_W$ at the eastern and western basin boundaries

$$T(y, z) = \frac{p_E(y, z) - p_W(y, z)}{\rho_f f},$$

(1)

where $f$ is the Coriolis parameter and $\rho_f$ the mean sea water density [e.g., Bingham and Hughes, 2008, 2009]. Integrating equation (1) between depth levels $z_1$ and $z_2$ yields the layer geostrophic AMOC volume transport from ocean bottom pressure data across the ocean basin [e.g., Kanzow et al., 2007; Bingham and Hughes, 2008]:

$$T(y) = \frac{1}{\rho_f f} \int_{z_1}^{z_2} p_E(y, z) - p_W(y, z) dz.$$  

(2)

Here we focus on OBP anomalies in the layer $z_1 = 3000$ m to $z_2 = 5000$ m. The focus is on this layer because (1) it has a sufficiently large horizontal extent that can be resolved by GRACE, (2) it is relatively far away from land to avoid hydrological signal leakage, and (3) it corresponds to the so-called Lower North Atlantic Deep Water (LNADW) layer that is observed with the RAPID-MOCHA array. As detailed in Hughes et al. [2013], the step-like bathymetry from 3000 to 5000 m along the western boundary implies that the bottom pressure gradients here will contribute most to the zonally averaged transport. More gently sloping topography would require additional information.

The southward LNADW return flow is itself well correlated with northward AMOC interannual variations (e.g., Figure 2 in McCarthy et al. [2012]) and can thus serve as a proxy of upper/northward AMOC variability. If a basin-mean or depth-averaged boundary pressure is removed from $p_W$ in equation (2), it is also possible to use only bottom pressure on the Western boundary, at least on interannual time scales (BH09). Here, however, we use the outermost points on both eastern (OBPE) and western (OBPW) boundaries to avoid issues arising from basin-mean bottom pressure variability, and land-hydrology signal leakage that can adversely affect GRACE OBP signals [e.g., Bentel et al., 2015]. Land hydrology signals can be an order of magnitude stronger than OBP variations, and due to GRACE’s limited resolution, coastal ocean signals can be contaminated with strong land signals [Chambers and Bonin, 2012; Bentel et al., 2015]. The effect of intervening topography from the Mid-Atlantic Ridge (MAR) is discussed below.

For our analysis, we use monthly OBP anomalies derived from the new JPL-RL05M mascon gravity fields. This GRACE solution is described in detail in Watkins et al. [2015], and we limit the data description here to a few points pertinent to the subsequent analysis. The JPL-RL05M uses a priori constraints in space and time to estimate global, monthly gravity fields in terms of equal area 3° spherical cap mass concentration functions (mascons). Compared to OBP anomalies derived from conventional GRACE spherical harmonic solutions [e.g., Chambers and Bonin, 2012], the mascon basis function coupled with the Bayesian constraints in the JPL-RL05M solution allows for improved spatial localization and S/N ratios, as well as better signal separation between land and ocean regions (see Watkins et al., 2015 for further details). As is standard practice, we remove solid Earth GIA trends [A et al., 2013], and restore the monthly dealiasing fields [Dobslaw et al., 2013] to obtain the full oceanographic OBP signal. Recently, an additional long-period correction for the pole tide (main components are at 12 and 14 month periods but with significant energy also at interannual and decadal periods) has been identified [Wahr et al., 2015], which we have included here as well. We use a modest Gaussian averaging filter of 50 km half width to smooth transitions across the mascon boundaries. OBP variations over steep bathymetry gradients cannot fully be resolved as a function of bottom depth since
the native resolution of the gravity solution is 3° but can nonetheless be detected if the signal is large enough. Comparisons with individual in situ bottom pressure recorders in the tropical North Atlantic have shown that JPL-RL05M resolves monthly to intraannual variations well, with RMS differences of 11–14 mm-H2O [Watkins et al., 2015]; 1 mm-H2O height corresponds to approximately 10 Pa. Note that this RMSD also contains noise and errors of the BPRs, and since the BPRs are point measurements, the RMSD is thus likely a conservative upper limit for the uncertainty of GRACE OBPs in this region.

We compare and evaluate our GRACE-derived AMOC component against data from the RAPID-MOCHA array [Smeed et al., 2014]. As GRACE’s gravity field variations are measured relative to a time-mean field, we can only infer OBP anomalies relative to the time-mean geostrophic transports. Thus, the following transport variations relative to the 2005–2012 mean circulation provide insight into the monthly to interannual variability of the mean transports. In all subsequent analysis, we have (1) subtracted a global mean ocean mass signal and monthly climatologies, (2) subtracted a linear least squares trend from all data points, and (3) applied a Lowess smoothing filter with a window width of 9 data points to focus on interannual OBP and AMOC variations. Note that as we detrend the OBP fields, Glacial Isostatic Adjustment (GIA) [e.g., A et al., 2013] effects and related uncertainties are effectively removed from OBPs here. After low-pass filtering, the monthly GRACE OBP values have an uncertainty of approximately 5 mm-H2O.

Figure 1. (a) Time series of eastern and western boundary bottom pressures (offset for clarity), averaged over 3000–5000 m at 26.5N. The variability on the western boundary is significantly larger (RMS: 6.1 mm) than on the eastern boundary (RMS: 3.5 mm), and the variance of the east-west difference (red line) is mostly explained by OBP variations at the western boundary (67%). Annual cycle and linear trend are subtracted, and 9 point “Lowess” filter has been applied (see section 2 for details); (b) zonal bathymetry profile, highlighting the western (solid black) and eastern (solid blue) basin boundaries between 3000 and 5000 m. The Mid-Atlantic Ridge eastern (dashed blue) and western (dashed black) flanks rise up to 3000 m, but OBP variations (dash lines in inset) along these flanks are nearly identical and thus effectively cancel out in the cross-basin gradients. The grey bar extends over 3° longitude, corresponding to the GRACE JPL-mascon horizontal resolution (see text for details). One mm-H2O OBP corresponds to approximately 10 Pa.
3. Results

We first evaluate GRACE-observed OBP variations at 26.5N, where the continuous in situ measurements of the RAPID-WATCH/MOCHA/WBTS array [e.g., Marotzke et al., 2002; McCarthy et al., 2015] provide a unique opportunity to compare our gravimetric transport time series with hydrographic estimates. Averaged over the 3000–5000 m depth interval (Figure 1a), the GRACE OBP anomalies from 2003 through 2014 reveal significantly higher variability on the western boundary (RMS: 6.1 mm-H2O), compared to the eastern side (3.9 mm-H2O). Therefore, the cross-basin east-west pressure difference is dominated by the western boundary signal (OBPW), which explains slightly more than 67% of the variance. These features are consistent with observations [Elipot et al., 2014; Smeed et al., 2014], theory [Hughes and de Cuevas, 2001], and ocean simulations (BH08). Possibly intervening topography (i.e., the MAR) should in principle be considered in the cross-basin transport computations. At and near 26.5N, the MAR rises to above 3000 m depth, thus fully intersecting with the cross-basin bottom pressure gradients below this depth (Figure 1b; “fully blocked” domain; BH08). However, we have evaluated OBPs on the eastern and western flanks of the MAR and found that they are approximately equal (Figure 1b inset; RMSD: 1.3 mm-H2O) and would thus effectively cancel out in the zonal pressure differential integration. Therefore, it is sufficient to use only the outermost eastern and western points across the basin section for the integration, consistent with Kanzow et al. [2007]. Note that by east-west differencing, we also eliminate any possible basin or global mean OBP mode (BH08) that is present in the GRACE fields.

The most prominent signals in boundary OBPs at 26.5N occur from late 2009 and last through early 2010 (Figure 1a), coinciding with a well-documented reduction of northward AMOC transports [e.g., Smeed et al., 2014]. The OBPw anomalies reach values of almost 20 mm-H2O. Again, this signal is largely absent in eastern boundary bottom pressures (Figure 1a). Looking at North Atlantic basin OBP anomalies from November 2009 through March 2010, prominent negative anomalies are detected north of about 40N, centered on the North Atlantic Current path that separates the subtropical from the subpolar gyre (Figure 2) and broadly consistent with the first EOF-expansion mode of GRACE bottom pressures between 30 and 60N [Piecuch and Ponte, 2014]. The mean nonseasonal bottom pressure anomalies range in amplitude from +50 mm-H2O on the shelf areas of the northern west Atlantic to about −35 mm-H2O in an area northeast of Bermuda. This
Figure 3. Meridional transport estimates from GRACE OBP anomalies on the eastern and western margin integrated over the 3000–5000 m depth layer at 26.5N, compared to the RAPID-MOCHA estimate of LNADW. The RMS difference between these two estimates is 1.2 Sv and the correlation is $R = 0.69$. The 1 sigma error of the GRACE-LNADW estimate is $\pm 1.1$ Sv.

Integration of the boundary OBPs yields the first satellite gravimetry-only estimate of the 3000–5000 m AMOC anomalies at 26.5N (Figure 3). Note that due to GRACE’s large footprint (Figure 1b), the 3000–5000 m pressure integration with equation (2) is effectively an average along the depth profile. Nonetheless, this gravimetric ocean transport estimate agrees very well with the in situ observations from the RAPID array. After monthly averaging of the RAPID-LNADW estimate and applying consistent low-pass filters (see section 2), we find a correlation of $R = 0.70$ and an RMS difference of 1.2 Sv between the two estimates for the time period of overlap May 2004 to March 2014 (Figure 3). Partitioning the integration (equation (2)) into the eastern and western part of the basin (i.e., east and west of the MAR; Figure 1b) slightly reduces the correlation ($R_{\text{MAR}} = 0.69$) and increases the RMSD (1.3 Sv), but the differences are not significant. The prominent 2009/2010 AMOC anomaly is effectively mirrored in the southward LNADW return flow (note that positive values correspond to a northward LNADW anomaly and hence a reduction in compensating southward return flow). Propagating the GRACE OBP errors mentioned above through the transport calculation (equation (2)), we estimate a 1 sigma error for a monthly LNADW value of approximately $\pm 1.1$ Sv. As such, differences between our GRACE and the RAPID LNADW estimate shown in Figure 3 comfortably fit within the uncertainties of each method. We also note that the RAPID-MOCHA AMOC algorithm imposes the constraint of zero net AMOC transport across latitudes [e.g., McCarthy et al., 2012], essentially implying that the cross-basin upper ocean northward transport (0–1000 m) must return in the southward return flow between 1000 and 5000 m depth. Intertidal northern Atlantic AMOC variations from RAPID have been found to be mainly mirrored in LNADW (3000–5000 m) [e.g., Kanzow et al., 2007; McCarthy et al., 2012; Smeed et al., 2014], and the LNADW estimate is thus highly correlated with the total northward AMOC transport. This so-called compensation of external transport had previously been evaluated and verified from bottom pressure recorders but due to drift was limited to subannual time scales [Kanzow et al., 2007; Frajka-Williams, personal communication, 2015]. Our direct bottom pressure integration from GRACE here can be interpreted to support (within the errors) the no-net-transport assumption even on interannual time scales. Possible net zonal mass transport variations arising from sources such as Bering Strait inflow and net freshwater input north of about 70N are likely of similar magnitude as the uncertainties and differences discussed above [e.g., Wunsch and Heimbach, 2013].

It is worth mentioning that the monthly mean of the GRACE dealiasing model (OMCT) that is restored during postprocessing [e.g., Chambers and Bonin, 2012] also shows similar correlations to the RAPID LNADW estimates as GRACE at 26.5N. To assess whether GRACE is providing skill at capturing the LNADW transport versus the background model OMCT, we processed a customized GRACE gravity solution for 1 year (2009) for which the monthly mean of OMCT is set equal to zero. The estimated monthly gravity anomalies are then completely independent of any monthly mean estimate of OBP in the background model. Over the study region, the
RMS differences between this new custom solution with a zero-mean background dealiasing model and the conventional solution are on the order of 5–10 mm-H2O, consistent with the expected uncertainty in the GRACE-derived gravity fields. This test indicates that GRACE is indeed skilful in capturing OBP variations related to the LNADW transport anomalies independent of any model. We also note that evaluations of GRACE-observed and OMCT-simulated OBP variations against BPRs have shown that GRACE observations generally

Figure 4. (a) Correlation between GRACE ocean bottom pressures and the GRACE estimate of Lower North Atlantic Deep Water transport at 26.5N (shown in Figure 3). The yellow contour line highlights areas where the correlation |R| ≥ 0.7; (b) Regression of LNADW against North Atlantic bottom pressure from GRACE observations to indicate AMOC-related OBP signal strength. Note that areas north of approximately 35N with apparent higher sensitivity are actually not highly correlated with LNADW variability at 26.5N. One mm-H2O OBP corresponds to approximately 10 Pa.
agree better to BPR measurements, in particular toward higher latitudes [Watkins et al., 2015]. This implies that GRACE-based transport estimates at other (higher) latitudes likely would also provide a better estimate of transport variability than a model-only estimate.

Given the completely different and independent nature of the gravimetric and in situ transport estimates at 26.5N, the level of agreement is very encouraging and highlights the potential of GRACE to evaluate the coherence of LNADW variations across nearby latitudes. To this end, we have evaluated the correlation of the LNADW at 26.5N at all grid points in the North Atlantic (Figure 4a). Correlation values exceeding \( R = 0.7 \) are generally limited to latitudes between 20N and 40N in the western basin, corresponding roughly to the meridional subtropical gyre extent. While the advection time scale for AMOC anomalies from 26N to 16N is about 2–3 years [Smeed et al., 2014], a lead-lag correlation analysis (not shown) has not revealed any obvious propagating features; the correlation values between GRACE OBPs and the corresponding 26.5 LNADW estimate are maximum at zero lag. However, the time series is still relatively short to allow a robust assessment of propagation on interannual time scales. Note also that the latitudinal extent of the interannual zero lag correlations in Figure 4a is broadly consistent with those of AMOC meridional coherence reported in ocean simulation studies, showing the highest correlation between approximately 20 and 40N [e.g., Bingham et al., 2007; Wunsch and Heimbach, 2013].

To evaluate the detectability of AMOC-related OBP variations, we also assessed the regression of the GRACE LNADW estimate at 26.5N across the North Atlantic (Figure 4b). This sensitivity (in units of mm-H2O Sv\(^{-1}\)) again confirms the dominance of the western boundary signals in terms of meridional transport variations. However, while the regression between OBP and LNADW yields a pattern that appears fairly continuously sensitive along the western side of the basin (amplitudes reach about 4–5 mm-H2O Sv\(^{-1}\)), higher correlation values \( |R| > 0.7 \) are much more spatially confined to around 26.5N (Figure 4a), where the cross-basin integration based on equation (2) can be carried out.

4. Conclusion

We have presented a first multiyear estimate of basin-wide Atlantic meridional interannual transport variations derived solely from satellite time-variable gravity measurements of the GRACE mission. Although this space-based measurement can currently not resolve spatial scales below approximately 300 km, we find very good agreement at 26.5N with our gravity-based estimate of the 3000–5000 m transport changes between 2004 and 2014 with those from the in situ RAPID array. The skill of the GRACE transport estimate arises mainly through a favorable combination of (1) low-land signal leakage for the deep LNADW layer away from land, (2) favorable step-like bathymetry for the 3000–5000 m layer [Hughes et al., 2013], and (3) the novel JPL-mascon GRACE processing that avoids empirical filters and maximizes the S/N ratio based on geophysical constraints [Watkins et al., 2015]. By removing the monthly mean ocean dealiasing information, we have also ascertained that the AMOC-related OBP signals reported here are indeed observed by the GRACE satellites, rather than being introduced through the background ocean model.

The previously documented LNADW transport reduction in 2009–2010 at 26.5N is well captured by the GRACE OBP integration. Consistent with earlier work [e.g., BH08; BH09; Elipot et al., 2014], the western boundary bottom pressure variability dominates, contributing approximately 2/3 of the LNADW variations. The viability and skill of GRACE-based transports estimates demonstrated here opens the door for spatially continuous measurements of AMOC transports, in particular on longer time scales. So-called external transports, which are imposed by a no-net-flow constraint or from BPRs, can be derived from GRACE and avoid drift problems inherent to in situ BPR recorders [e.g., McCarthy et al., 2012]. Gravimetric OBP observations are not affected by instrument drift, and basin-scale or global-mean OBP signals will be removed by taking cross-basin differences. Nonetheless, we have detrended all time series here due to uncertainty in the GRACE trend corrections for GIA and land leakage, which require a detailed trend-error evaluation of the GRACE and RAPID/MOCHA signals beyond the scope of this paper. We intend to pursue this topic in future studies.

The direct integration of zonal bottom pressure gradients as demonstrated here is currently limited by the ability of GRACE to measure them below spatial scales of about 300 km. A closer inspection of Figure 3 also indicates that the large LNADW anomaly in 2009/2010 significantly contributed to the good agreement between GRACE and RAPID measurements. Completely removing the 2009–2010 period reduces the correlation to \( R = 0.45 \). Nonetheless, we find that the sensitivity of the GRACE satellites is sufficient to record interannual transport anomalies in AMOC components, as long as they are sufficiently large (approximately
Acknowledgments

We thank Elanor Frajka-Williams for her helpful discussions about the RAPID-MOCHA measurements, and Don Chambers and an anonymous reviewer for their very constructive comments. This work represents one phase of research carried out at the Jet Propulsion Laboratory/California Institute of Technology, under a contract with the National Aeronautics and Space Administration. The JPL-RLO5M GRACE solutions are available via the Physical Oceanography Distributed Active Archive Center (PODAAC) as well as the GRACE Tellus websites (www.grace.jpl.nasa.gov). RAPID is funded by the National Environmental Research Council in the UK and the National Science Foundation and Oceanic and Atmospheric Administration in the United States. Data are freely available from http://rapid.ac.uk.

The Editor thanks Don Chambers and an anonymous reviewer for their assistance in evaluating this paper.

References


Bentaleb, E., F. W. Landeier, and C. Boening (2015), Monitoring Atlantic overturning circulation variability with GRACE (2015), Monitoring MOCHA measurements, and Don Chambers and an anonymous reviewer for their very constructive comments. This work represents one phase of research carried out at the Jet Propulsion Laboratory/California Institute of Technology, under a contract with the National Aeronautics and Space Administration. The JPL-RLO5M GRACE solutions are available via the Physical Oceanography Distributed Active Archive Center (PODAAC) as well as the GRACE Tellus websites (www.grace.jpl.nasa.gov). RAPID is funded by the Natural Environment Research Council in the UK and the National Science Foundation and Oceanic and Atmospheric Administration in the United States. Data are freely available from http://rapid.ac.uk.

The Editor thanks Don Chambers and an anonymous reviewer for their assistance in evaluating this paper.

References


Bentaleb, E., F. W. Landeier, and C. Boening (2015), Monitoring Atlantic overturning circulation variability with GRACE (2015), Monitoring MOCHA measurements, and Don Chambers and an anonymous reviewer for their very constructive comments. This work represents one phase of research carried out at the Jet Propulsion Laboratory/California Institute of Technology, under a contract with the National Aeronautics and Space Administration. The JPL-RLO5M GRACE solutions are available via the Physical Oceanography Distributed Active Archive Center (PODAAC) as well as the GRACE Tellus websites (www.grace.jpl.nasa.gov). RAPID is funded by the Natural Environment Research Council in the UK and the National Science Foundation and Oceanic and Atmospheric Administration in the United States. Data are freely available from http://rapid.ac.uk.

The Editor thanks Don Chambers and an anonymous reviewer for their assistance in evaluating this paper.