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Can in situ floats and satellite altimeters detect long-term changes in Atlantic Ocean overturning?

Josh K. Willis[1]

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[1] Global warming has been predicted to slow the Atlantic Meridional Overturning Circulation (AMOC), resulting in significant regional climate impacts across the North Atlantic and beyond. Here, satellite observations of sea surface height (SSH) along with temperature, salinity and velocity from profiling floats are used to estimate changes in the northward-flowing, upper limb of the AMOC at latitudes around 41°N. The 2004 through 2006 mean overturning is found to be 15.5 ± 2.4 Sv (10^6 m^3/s) with somewhat smaller seasonal and interannual variability than at lower latitudes. There is no significant trend in overturning strength between 2002 and 2009. Altimeter data, however, suggest an increase of 2.6 Sv since 1993, consistent with North Atlantic warming during this same period. Despite significant seasonal to interannual fluctuations, these observations demonstrate that substantial slowing of the AMOC did not occur during the past 7 years and is unlikely to have occurred in the past 2 decades. **Citation:** Willis, J. K. (2010), Can in situ floats and satellite altimeters detect long-term changes in Atlantic Ocean overturning?, *Geophys. Res. Lett.*, *37*, L06602, doi:10.1029/2010GL042372.

1. Introduction

[2] The AMOC carries warm surface waters into the North Atlantic Ocean, where they cool, sink and spread southward at depth. The resulting heat transport makes the Atlantic unique in that it is the only ocean that carries heat northward throughout both hemispheres. Popularized by the movie *The Day After Tomorrow*, the past and future of the AMOC as well as its role in global and regional climate has been a source of considerable scientific debate [*Schiermeier*, 2006]. For instance, it has been hypothesized to play a pivotal role in past climate change events when the North Atlantic was flooded by freshwater from melting ice. The introduction of light, fresh water should inhibit sinking, slow the overturning and reduce the northward transport of heat. Evidence suggests that such slowdowns drove rapid cooling events in the North Atlantic around the time of the end of the last ice age [*McManus et al.*, 2004].

[3] Although a complete shutdown in the next century appears unlikely, surface warming and changes in the hydrologic cycle from global warming are predicted to slow the AMOC in the future [*Meehl et al.*, 2007]. Although acceleration in the rate of ice loss from Greenland has already been well documented [*Rignot and Kanagaratnam*, 2006; *Wouters et al.*, 2008], model results suggest that the current ice loss rates are too small to slow the AMOC [*Hu et al.*, 2009]. Nevertheless, campaigns to continuously monitor its strength using moorings and ship-based hydrographic sections are already underway [*Cunningham et al.*, 2007].

[4] Early estimates of changes in the AMOC based on five hydrographic transects at 26.5°N suggested that a 30% slowing had already occurred between 1957 and 2004 [*Bryden et al.*, 2005]. More recently, however, mooring data have shown that large seasonal to interannual fluctuations occur in AMOC strength at that latitude, and the trend computed from the five original transects could not be considered statistically significant [*Cunningham et al.*, 2007]. Furthermore, decadal changes in the freshwater content of the subpolar North Atlantic [*Curry and Mauritzen*, 2005; *Boyer et al.*, 2007; *Sarafanov et al.*, 2008] have also been related to decadal changes in overturning strength [*Biastoch et al.*, 2008].

[5] Despite its importance, direct observations of overturning variability have so far been confined to the moorings and occasional hydrographic sections at 26.5°N. In the present study, data from satellite observations of SSH are combined with data from the Argo array of profiling floats to estimate the magnitude and variability of the AMOC near 41°N. The remainder of the manuscript is organized as follows. Section 2 discusses the data and techniques used to estimate the time varying absolute dynamic height field. Section 3 discusses the AMOC estimate and its errors. Results are presented in Section 4 and Discussion and Conclusions in Section 5.

2. Argo, Altimetry and Subsurface Velocity Estimates

[6] The method for estimating transports from Argo and altimeter data is based on the technique developed by *Willis and Fu* [2008], where it was tested extensively against North Atlantic transports based on a wide variety of observational data. Subsurface float data from the Argo array provide direct observations of time-integrated subsurface velocity by drifting at a fixed depth of 1000 m, 1500 m or 2000 m, for a period of several days. At the end of this drift period, floats dive to a depth of approximately 2000 m before ascending to the surface while measuring temperature, salinity and pressure. Using the subsurface drift observations to provide an estimate of dynamic height at 1000 m and the hydrographic profiles to estimate geostrophic shear, a three dimensional estimate of absolute dynamic height was computed from the surface to a depth of two kilometers (Figure 1).

[7] The high spatial and temporal sampling of the altimeter and the strong correlation between SSH and subsurface changes allow the altimeter data to be used to reduce sam-

[1]Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, USA.

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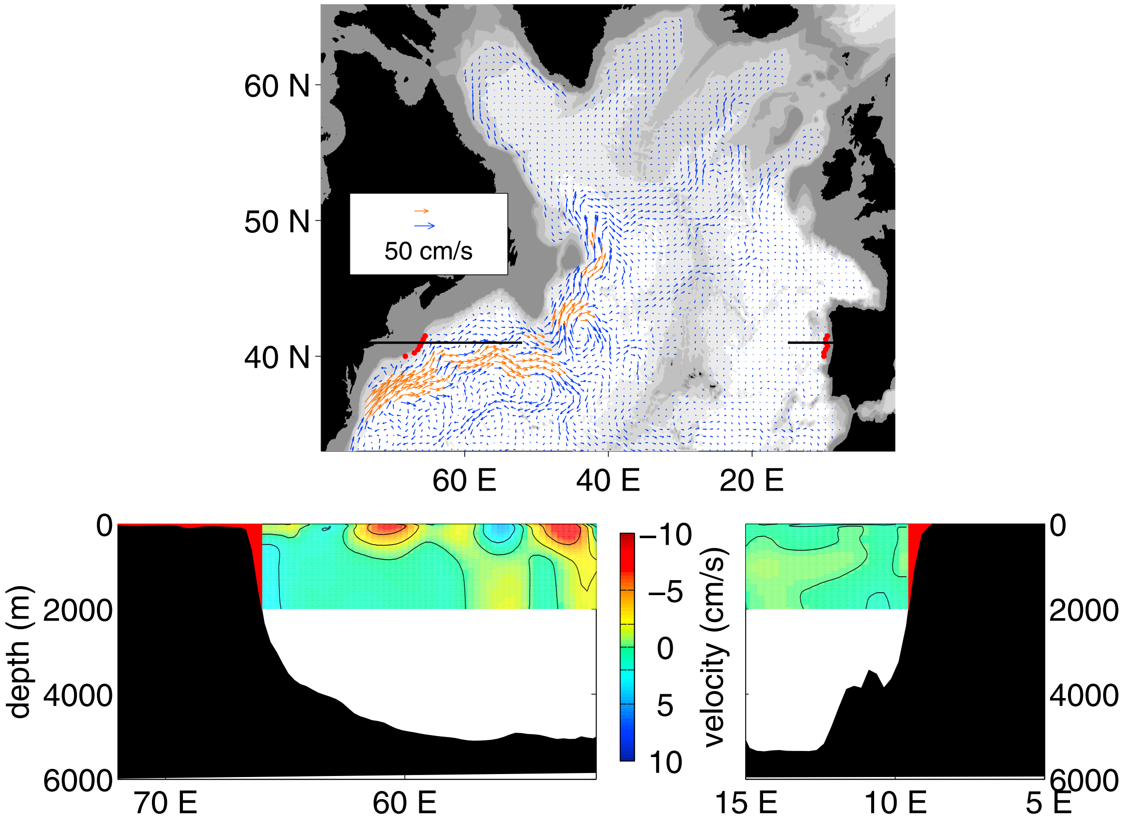

Figure 1. (top) 2004 through 2006 time-mean of surface geostrophic currents based on based on absolute dynamic height estimated using Argo and altimeter data in the study region. Arrows are shown every 1° of longitude and ½° of latitude for clarity. Orange vectors have been scaled down by a factor of 2 and generally lie within the Gulf Stream. The red dots show the basin boundaries marked by the 2000 m isobath between 40–41.5°N, the latitudes at which the overturning is estimated. From light to dark, shading indicates bathymetry <4000 m, <3000 m, <2000 m, <1000 m, and < 500 m. (bottom) Black lines in the top panel indicate transects at 41°N along which northward velocity is plotted. Colors indicate velocity and contours are as indicated in the color scale. The red areas mark the shallow boundary regions that are not sampled by Argo.

pling error in observations of temperature, salinity and subsurface velocity. *Willis and Fu* [2008] used altimeter data to reduce errors in time averaged estimates of the circulation. In the present study, altimeter data is further used to reduce errors in time-varying estimates.

[8] Maps of the time-varying 1000 m dynamic height and upper 2000 m density field were computed from 35° to 66°N for each month using data in overlapping three-month time windows (see auxiliary material for details on how maps of density and 1000 m dynamic height variability were computed).[1] The 3-dimensional absolute dynamic height field was then computed using the thermal wind equations on each 3-month time average of the density and 1000 m dynamic height fields.

3. Estimating the AMOC Transport and Its Error

[9] Because they dive to a depth of approximately 2000 meters to obtain temperature and salinity profiles, the Argo floats do not sample the currents in shallow regions like the continental shelf or the upper part of the continental slope. Argo-based estimates of geostrophic velocity are therefore restricted to regions where the water is deeper than 2000 m (Figure 1). This precludes using Argo to estimate the AMOC at latitudes where significant meridional transport lies in shallow regions. For instance, at 33°N, much of the Gulf Stream transport sits on the continental shelf and slope (Figure 1). Because Argo cannot sample these shallow regions, the Argo-based estimate misses much of this northward transport and gives a basin-integrated transport that is much too low (Figure S1).

[10] To determine whether any latitudes exists where lack of sampling in shallow regions does not cause large errors in basin-integrated transport, the ECCO2 (Estimating the Circulation and Climate of the Ocean, Phase II) high-resolution, global ocean general circulation model was used [*Menemenlis et al.*, 2008] (also see http://ecco2.jpl.nasa.gov). The model also appears to have a realistic Gulf Stream and AMOC (Figure S2, see auxiliary material for more details).

[11] To estimate the error caused by lack of data in the shallow regions, AMOC transport in three-month averaged ECCO2 fields was computed by calculating the absolute dynamic height at the boundaries determined by the 2000 m isobaths and subtracting them. The geostrophic model transports were integrated between the surface and 958 m. Model Ekman transports were computed from the wind stress fields used to force the model and were added to the geostrophic transports. This was compared with basin-

[1] Auxiliary materials are available in the HTML. doi:10.1029/2010GL042372.

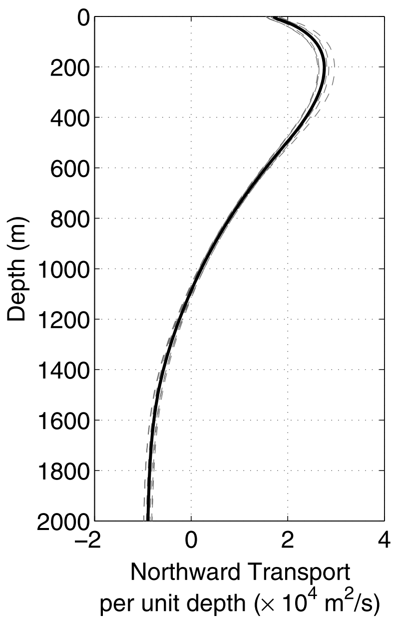

Figure 2. Geostrophic transport per unit depth at latitudes between 40–41.5°N at each ¼° latitude grid (thin dashed lines) from the 2004 through 2006 time-average circulation. The mean over these latitudes is shown as a thick black line.

integrated transport computed directly from the complete model velocity fields (Figure S1). North of the boundary current separation and south of the Grand Banks, a narrow band of latitudes was found where lack of shallow-water velocities resulted in RMS errors of only 1.1 Sv (Figure S1). The mean difference between 40–41.5°N, averaged over the 12 years of the model run was −0.32 Sv. This is significantly smaller than sampling errors in the observations.

[12] The error in the geostrophic transport was estimated in two parts. First the sampling error in the estimate of the circulation was computed at the location of the east and west boundary at 41°N defined by the 2000 m isobath. Geostrophic transport was computed as the difference between dynamic heights at these two locations. The error was therefore computed as the root-sum-square of the error in dynamic height at the east and west boundaries as determined by the objective map. These error estimates were found by *Willis and Fu* [2008] to be robust measures of the sampling error. Although the estimates of overturning are averaged over latitudes from 40–41.5°N, the error estimates are not reduced to reflect this averaging, as the covariance length scales are comparable to or larger than these meridional distances (see Table S1). The total error in the AMOC transport was then computed for each three-month averaged estimate as: $E_{total} = \mathrm{sqrt}(E^2_{geostrophic\ transport} + (1.1\ \mathrm{Sv})^2)$, where the 1.1 Sv represents the error due to lack of sampling in the shallow regions. The RMS value of E_{total} over the 2002–2009 time series is 2.2 Sv.

[13] The transport for the 2004 through 2006 time-averaged circulation from *Willis and Fu* [2008] is shown in Figure 2. The vertical structure of the transport is similar to that of other latitudes [*Bryden et al.*, 2005; *Cunningham et al.*, 2007] with northward transport in the upper layers, a zero near 1130 m, and southward return flow below. Here the amplitude of the AMOC is defined to be the basin-integrated transport from the surface to 1130 m. Although the exact depth of the zero transport varies by 50 m or so, integrating to a variable depth produces a time series that is indistinguishable from the 0 to 1130 m integration. For this period, the geostrophic AMOC transport was 17.6 Sv and the Ekman transport was 2.1 Sv to the south. This gives a net overturning of 15.5 ± 2.4 Sv during this period. Here the 2.4 Sv represents the RMS variability of the overturning (including Ekman transport) of the three month averages about the 3-year mean. This is somewhat smaller than the 18.7 Sv mean value at 26.5°N as observed by *Cunningham et al.* [2007].

[14] It is assumed that all of the northward transport in the upper 1130 m is returned at depth. The ECCO2 model, which includes the Arctic, suggests that this is a reasonable assumption. In the model, the net transports above and below 958 m have a correlation of −0.985 and differ by only 1.4 ± 0.4 Sv in the time-mean, which represents the model flow through the Bering Strait. Furthermore, the 0.4 Sv RMS variability of this flow is much smaller than the variability in the AMOC as discussed below. Finally, in the section at 26.5°N, *Kanzow et al.* [2007] showed that compensation occurred in the layer below 1050 m on time scales longer than 15-days.

4. Results

[15] As described in Section 3, the time-varying maps of dynamic height were used to compute a time series of the geostrophic component of the overturning transport (Figure 3). Although use of the altimeter data reduces errors in the time-varying estimate of geostrophic transport, using only the time-mean and seasonal cycle as the initial guess gives very similar results (Figure 3).

[16] The temporal variability of the AMOC at 41°N is significantly smaller than the variability at 26.5°N. To compare, 3 years of data from the RAPID array were retrieved from http://www.noc.soton.ac.uk/rapidmoc. The RMS variability about the time-mean at 26.5°N is 2.4 Sv

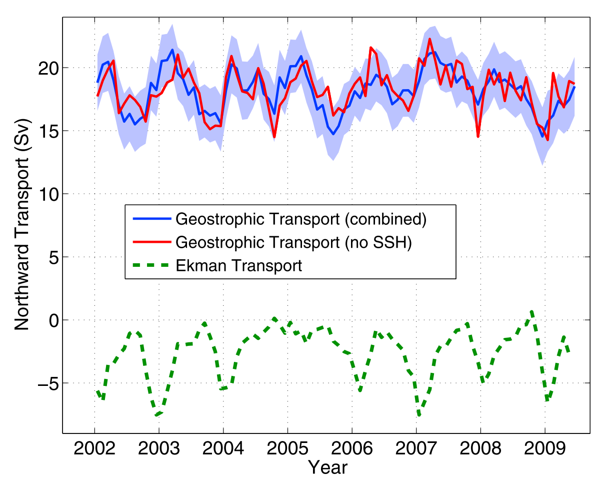

Figure 3. Geostrophic transport in the upper 1000 m based on a combination of Argo and altimeter data (blue line). Shading indicates error bounds. The red line shows another estimate of the geostrophic transport based on Argo data alone, without using the altimeter data to reduce sampling error. The dashed green line shows the basin integrated Ekman transport over the same period.

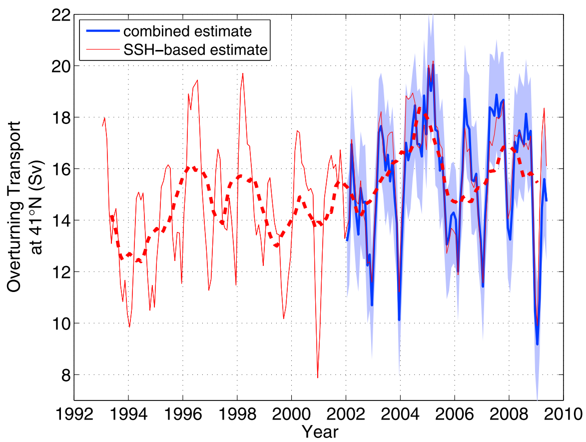

Figure 4. Variability in the AMOC at 41°N computed as the sum of the Ekman transport and the geostrophic transport between the surface and 1130 m using a combination of Argo and altimeter data (blue line). Shading indicates error bounds. The thin red line shows an estimate of the AMOC where the geostrophic transport is computed using only the altimeter data and regression coefficients. The dashed red line shows a one-year running mean of the SSH-based estimate.

for the three-month averaged geostrophic transport between 2004 and 2007. Furthermore, it has a strong seasonal cycle with an amplitude of 3 to 4 Sv. At 41°N, the RMS variability is 1.4 Sv, and the seasonal cycle is smaller and more irregular than that at 26.5°N.

[17] To estimate the total overturning at 41°N, the Ekman transport is added to the geostrophic transport in the upper 1130 m (Figure 3). Ekman transports were computed using wind stress products from the NCEP/NCAR Reanalysis monthly mean derived datasets [*Kalnay et al.*, 1996]. The Ekman component has considerably more seasonal to interannual variability, giving the total AMOC an RMS variability of 2.4 Sv at this latitude. This is still smaller than the AMOC variability at 26.5°N, which is 3.1 Sv for the 3-month average time series including both the geostrophic and Ekman components. It is also consistent with the range of values computed by *Lumpkin et al.* [2008, Table 3] using inverse methods on 5 transects at 48°N along with mooring data and surface flux estimates.

[18] The estimate of overturning variability shown in Figure 3 was computed using altimeter data and regression coefficients to provide an initial guess for both the density field and the 1000 m circulation. These initial guess fields, or altimeter-based estimates of the geostrophic circulation, can also be computed for the entire period of the altimeter record. Adding Ekman transport provides an estimate of the AMOC from 1993 through 2009 (Figure 4). A small trend is present in the altimeter-based estimate, with the overturning increasing by 2.4 ± 1.6 Sv over the 16-year period. Here, the 1.6 Sv uncertainty refers to the 95% confidence interval on the linear fit, suggesting that the trend is significant relative to the seasonal and interannual fluctuations. This is not a realistic observational error, however, and should be considered with some caution as discussed below.

[19] A one-year low-pass filter shows interannual variations of several Sv, particularly in the latter part of the record. Beginning in early 2002, the overturning began to increase rapidly by about 4 Sv. It peaked in late 2004 before slowing again by about 3 Sv during 2005. In the low-pass filtered estimate, the Ekman transport remains an important but smaller component of the AMOC variability, accounting for about 35% of the interannual variations.

5. Discussion and Conclusions

[20] These observations suggest that the time-averaged overturning at 41°N is 2 to 3 Sv smaller than in the mid-latitudes [*Cunningham et al.*, 2007]. Seasonal to interannual variations in the overturning also appear to be significantly smaller than at lower latitudes, which may prove advantageous for detecting decadal changes in the strength of the AMOC.

[21] No significant trend was detected in AMOC transport between 2002 and 2009. The altimeter record, however, suggests a slight strengthening since 1992. Although agreement with Argo-based observations during the latter years is encouraging, some care is required for interpreting the altimeter-based estimate over the entire 16-year period. Regression coefficients between altimeter observations and individual profiles are 0.5 to 0.6, while regression coefficients with subsurface displacement data are closer to 0.3 [*Willis and Fu*, 2008]. Nevertheless, on long enough time scales, temperature and salinity anomalies advected into the domain could introduce significant error into these coefficients. This underscores the need for ongoing hydrographic observations such as those obtained by the Argo floats as well as the more traditional ship-based surveys.

[22] Based on coupled climate model runs, *Knight et al.* [2005] suggested a connection between surface temperature of the North Atlantic and AMOC strength. Despite uncertainty in the early part of the 16-year record, the increase in AMOC strength during the 1990s is consistent with decadal warming in the North Atlantic relative to the South Atlantic during the 1980s and 1990s. The decadal variations in AMOC strength may also be consistent with decadal changes in the temperature and salinity of the subpolar gyre [*Sarafanov et al.* 2008; *Boyer et al.*, 2007; *Curry and Mauritzen*, 2005] but further work is needed to determine the dynamical link between these property changes and their relation to changes in the AMOC [*Biastoch et al.*, 2008].

[23] The estimate presented here is the longest direct observation of overturning variability to date, and the only one at high latitude. Given its potential consequences for both regional and global climate change, monitoring the AMOC remains an important observational priority. The technique presented here provides estimates of AMOC strength with errors of roughly 2 Sv in 3-month average estimates. Whether driven by global warming or part of its natural variability, decadal changes in the AMOC that are larger than 1 to 2 Sv should be readily apparent in the Argo and altimeter observations. Global observing systems such as these provide an important, high-latitude complement to the moorings and ship-based hydrographic sections that are currently being used to monitor the AMOC at lower latitudes.

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J. K. Willis, Jet Propulsion Laboratory, California Institute of Technology, 4800 Oak Grove Dr., Pasadena, CA 91109, USA.