Observed interannual variability of the Atlantic meridional overturning circulation at 26.5°N

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[1] The Atlantic meridional overturning circulation (MOC) plays a critical role in the climate system and is responsible for much of the heat transported by the ocean. A mooring array, nominally at 26°N between the Bahamas and the Canary Islands, deployed in Apr 2004 provides continuous measurements of the strength and variability of this circulation. With seven full years of measurements, we now examine the interannual variability of the MOC. While earlier results highlighted substantial seasonal and shorter timescale variability, there had not been significant interannual variability. The mean MOC from 1 Apr 2004 to the 31 March 2009 was 18.5 Sv with the annual means having a standard deviation of only 1.0 Sv. From 1 April 2009 to 31 March 2010, the annually averaged MOC strength was just 12.8 Sv, representing a 30% decline. This downturn persisted from early 2009 to mid-2010. We show that the cause of the decline was not only an anomalous wind-driven event from Dec 2009–Mar 2010 but also a strengthening of the geostrophic flow. In particular, the southward flow in the top 1100 m intensified, while the deep southward return transport—particularly in the deepest layer from 3000–5000 m—weakened. This rebalancing of the transport from the deep overturning to the upper gyre has implications for the heat transported by the Atlantic. **Citation:** McCarthy, G., E. Frajka-Williams, W. E. Johns, M. O. Baringer, C. S. Meinen, H. L. Bryden, D. Rayner, A. Duchez, C. Roberts, and S. A. Cunningham (2012), Observed interannual variability of the Atlantic meridional overturning circulation at 26.5°N, *Geophys. Res. Lett.*, *39*, L19609, doi:10.1029/2012GL052933.

1. Introduction

[2] The Atlantic meridional overturning circulation (MOC) is responsible for almost 90% of the net meridional ocean heat flux of 1.3 PW near 26°N [*Johns et al.*, 2011]. Ultimately this heat transport is responsible for the mild maritime climate of north western Europe [*Rhines et al.*, 2008], where surface air temperatures are 5–10°C warmer than the global zonal average at those latitudes [*Rahmstorf and Ganopolski*, 1999]. It is expected, based on the response of coupled climate models to increasing atmospheric CO_2 concentrations, that the MOC will slow by 25% over the next few decades [*Bindoff et al.*, 2007].

[3] Since 2004, we have been measuring the strength and vertical structure of the MOC at 26°N using a transatlantic array known as the RAPID-WATCH/MOCHA/WBTS array (hereafter the 26°N array) [*Rayner et al.*, 2011]. These long-term sustained measurements provide fundamental observations for understanding the MOC and its response to forcing on climate relevant timescales, a baseline for measuring future MOC change and initial conditions for climate forecasts.

[4] From 1 Apr 2004 to 31 Mar 2009, the mean MOC strength was 18.5 Sv. However beginning in early 2009 and lasting until mid-2010, the MOC slowed below its mean value. The annual mean from 1 April 2009 to 31 March 2010 was 12.8 Sv. In this paper, we show that this change is partly due to extreme Ekman transport from Dec 2009–Mar 2010 but mainly due to a longer timescale strengthening of southward gyre transport above 1100 m. This strengthening of the southward thermocline flow was accompanied by decreased southward flow in the deep limb of the overturning below 3000 m: in effect, a rebalancing of the 26°N circulation from overturning to gyre circulation.

2. Methods

[5] The 26°N array combines measurements of the Gulf Stream in the Florida Straits [*Baringer and Larsen*, 2001], Ekman transports calculated from Cross-Calibrated Multiplatform winds [*Atlas et al.*, 2011] (see also auxiliary material) and mid-ocean transports from the 26°N mooring array.[1] The mooring array consists of two parts. From the Bahamas to 20 km offshore (76.75°W), current meter moorings make direct estimates of the flow [*Johns et al.*, 2008]. East of 76.75°W, a dynamic height array defines the mid-ocean geostrophic flow to the eastern boundary, south of the Canary Islands. These transports combined are referred to as the internal transport. Geostrophic calculations require a level of no or known motion, which, for the MOC calculation, is determined by the constraint of zero net mass transport across the section. The additional transport necessary to satisfy this constraint of zero mass transport across the section is referred to as the external transport. While the external transports are defined to satisfy an imposed constraint, they have been independently validated using *in situ* bottom

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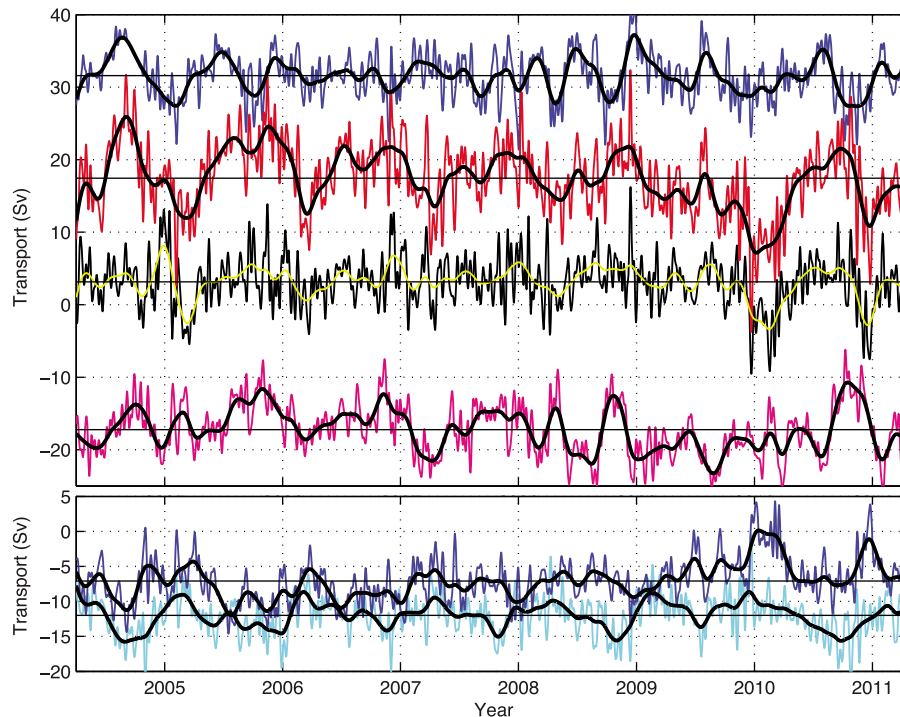
Figure 1. (top) Ten-day (colors) and three month low-pass (black or yellow for Ekman) timeseries of Gulf Stream transport (blue), Ekman transport (black), upper mid-ocean transport (magenta), and overturning transport (red) for the period 1 April 2004 to 22 April 2011. Horizontal black lines are timeseries means. A Butterworth filter, passed forward and backwards to avoid phase shifting, was used on the data. (bottom) Layer transports for upper NADW in the depth range 1100 to 3000 m (light blue), and lower NADW in the depth range 3000 to 5000 m (dark blue). The sum of the upper and lower NADW Transports is the lower limb of the overturning and has a correlation of $r = -0.9958$ with the MOC. Positive transports correspond to northward flow.

pressure data by *Kanzow et al.* [2007]. The internal and external transports are combined to give a zonally integrated mid-ocean transport profile as a function of depth. The mid-ocean transport profiles are further combined with the Gulf Stream and Ekman transport profiles, then vertically integrated to give the overturning transport streamfunction. The MOC is defined as the maximum of this streamfunction and the depth of the MOC is the depth of this maximum: predominantly at 1100 m (see Figure S1 in the auxiliary material for further details). We define the mid-ocean transports down to the depth of the maximum streamfunction as the upper mid-ocean transports (UMO). Hence,

$$UMO + Gulf\ Stream + Ekman = MOC.$$

[6] Due to extreme events in late December 2009 and again in early January 2010, some time periods have no net northward transport of upper water (see Figure S1 in the auxiliary material for details on the MOC definition for these periods).

[7] Annual averages are calculated from 1 April of a year through 31 March of the following year and are referred to as, for example, 09/10 for the period from 1 Apr 2009–31 Mar 2010 or 2004–2008 for the period 1 Apr 2004–31 Mar 2009.

[8] As this paper primarily deals with interannual changes, standard deviations are calculated from the time series of the annual averaged values as stated in the text. The reference period is taken to be 2004–2008, so changes refer to the difference from the reference period to the 09/10 year.

Results do not differ substantially if the reference period is changed to include the 10/11 year nor if the whole timeseries is used as a reference period.

3. Results

[9] For the first five years the annual mean MOC varied little from a mean of 18.5 Sv (17.8, 19.9, 19.3, 18.0, 17.5 Sv), with no significant interannual variability. Annual means of its constituent components: Gulf Stream, Ekman and upper mid-ocean transports also vary little over this period (Figure 1 and Table 1). Beginning in early 2009 the MOC strength dropped below its mean value, remaining low until mid-2010. The annual mean MOC strength from 1 Apr 2009 through 31 Mar 2010 was only 12.8 Sv, reaching a minimum on 22 Dec 2009 of -3.7 Sv. While the negative MOC is remarkable, the longer term changes represent a reduction of more than 30% that was sustained for over a year. This change of 5.7 Sv is more than five times the standard deviation of the annual averages over the preceding five years.

[10] The reduction in MOC strength is driven by changes in its constituent components: a 1.1 Sv reduction in the northward Gulf Stream, a 1.7 Sv reduction in the northward Ekman transports and a 2.9 Sv increase in the southward upper mid-ocean circulation (Table 1). The causes of a slight reduction in the Gulf Stream have been identified as short timescale processes and will not be discussed further here. Ekman changes were anomalously negative but not as large

Table 1. Values of Meridional Transport of Components of the MOC in Units of Sverdrups[a]

| | Annual Mean | | | | | | | Mean ± sd | Change |
	2004	2005	2006	2007	2008	2009	2010	2004–2008	2009–Ref
	Shallow Components								
Ekman transport	3.0	3.4	3.8	3.7	3.7	1.8	2.7	3.5 ± 0.3	−1.7
Upper mid-ocean transport	−17.0	−15.5	−16.1	−17.1	−18.3	−19.7	−16.8	−16.8 ± 1.1	−2.9
Gulf Stream	31.8	32.0	31.6	31.6	32.2	30.7	31.0	31.8 ± 0.3	−1.1
MOC	17.8	19.9	19.3	18.0	17.5	12.8	16.9	18.5 ± 1.0	−5.7
	Deep Layers								
UNADW (1100–3000 m)	−12.2	−12.5	−12.0	−11.8	−11.8	−10.9	−13.0	−12.0 ± 0.3	1.1
LNADW (3000–5000 m)	−7.3	−8.8	−8.6	−7.6	−7.3	−4.0	−5.9	−7.9 ± 0.7	3.9
Deeper than 5,000 m	2.0	1.7	1.6	1.8	1.9	2.5	2.4	1.8 ± 0.1	0.7

[a]Positive values are northward. Each annual mean is computed from 1 Apr–31 Mar inclusive, and the eighth column is the mean and standard deviation over the reference of the first five years of annual means while the far right column is the difference of the reference period from the 09/10 year. The shallow components sum to the value of the MOC. Deep layers are defined in depth classes and roughly sum to negative MOC.

or long-lived as the upper mid-ocean changes. Thus the substantially weakened MOC in 09/10 is mainly due to an increase in the southward upper mid-ocean circulation.

3.1. Changes in the Ekman Transport

[11] At 26°N the zonal average winds are predominantly easterly giving rise to a mean northward Ekman transport of 3.5 Sv (Figure 1 and Table 1). In the winters of 04/05, 09/10 and 10/11 the zonal average winds reversed to the westerly direction for a time, resulting in a southward Ekman transport. While the 09/10 and 10/11 negative Ekman transports are related to winters with extreme negative NAO indices [*Wang et al.*, 2010], the 04/05 negative Ekman transport is not. All three anomalous Ekman events correspond to low MOC periods. However, the anomalously low Ekman values in winters 04/05 and 10/11 do not result in similar longer term downturns in the MOC as is observed in 09/10.

[12] To isolate the impact of the winds on the MOC, we replace the time-varying Ekman transport with a time-mean transport, and calculate $MOC_{\langle ek \rangle}$, its components, and layer transports, where the subscript $_{\langle ek \rangle}$ refers to the time-mean Ekman calculation. The mean $MOC_{\langle ek \rangle}$ in 09/10 is 4.3 Sv lower than the mean from 2004–2008, rather than the 5.7 Sv lower for the complete MOC estimate. Thus the anomalous winds contribute a 1.4 Sv reduction in the annual mean MOC, while the $GS_{\langle ek \rangle}$ and $UMO_{\langle ek \rangle}$ changes contribute 4.3 Sv. This is larger than the individual changes of the UMO and GS, 2.9 Sv and 1.1 Sv respectively, because some of the compensation to the Ekman transport is applied to the UMO transport via the mass balance constraint. The $UMO_{\langle ek \rangle}$ in 09/10 was 3.1 Sv stronger southward than in the 2004–08 reference period.

3.2. Balance Between the Upper and Deep Transports

[13] The southward flow in the full-depth mid-ocean transports from the Bahamas to the Canary Islands exactly balances the northward Gulf Stream transport when using a time-mean Ekman transport. This is necessary to balance mass across the section. Consider the decomposition of this full depth mid-ocean transport into upper and deep mid-ocean transport–above and below 1100 m respectively. If the Gulf Stream is solely balanced by the upper mid-ocean transport, then the circulation is that of an idealised subtropical gyre. If the Gulf Stream is solely balanced by the deep mid-ocean transport, then the circulation is an idealised overturning circulation. Since the Gulf Stream varies little

on interannual timescales, the primary balance in the large scale circulation at 26°N is between the upper and deep mid-ocean transports. Figure 2 shows the 09/10 downturn in the MOC driven by increased southward transport in the upper mid-ocean as well as the anomalous southward Ekman transports, as discussed in the previous section. The deep mid-ocean transports–the upper and lower North Atlantic Deep Water (NADW)–are less southwards as the balance shifts from overturning to gyre circulation.

[14] The relative strengths of the upper gyre and deep overturning transports are critical for estimates of heat transport. In particular, *Johns et al.* [2011] showed that while the gyre transport is of similar strength to the overturning, it carries far less heat per unit transport. This is as, when the warm northward Gulf Stream is balanced by southward flow in the warm waters of the gyre, the temperature difference is relatively small leading to lower heat transport. In contrast, when the Gulf Stream is balanced by southward flow in the cold, deep waters of the overturning,
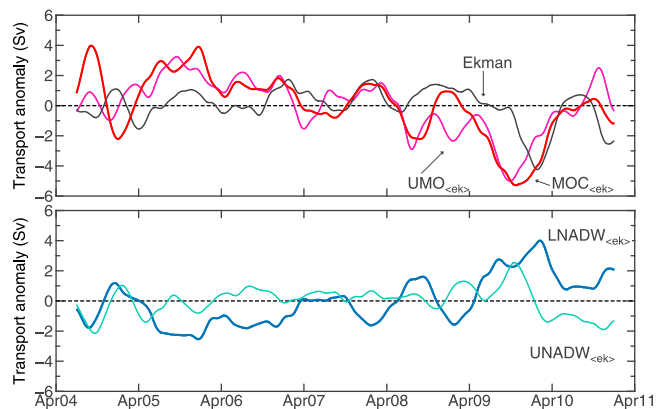
Figure 2. (top) Transport anomaly time series after removing the average seasonal cycle (calculated over the full timeseries) and smoothing with a 180-day low-pass Tukey filter. Components include Ekman (black), upper mid-ocean with fixed Ekman (magenta) and overturning with fixed Ekman (red). Transports in units of Sverdrups. (bottom) Transport time series with the average seasonal cycle removed of the upper NADW (1100–3000 m, cyan) and lower NADW (3000–5000 m, dark blue), calculated with fixed Ekman. These are the main water masses of the deep mid-ocean transports.
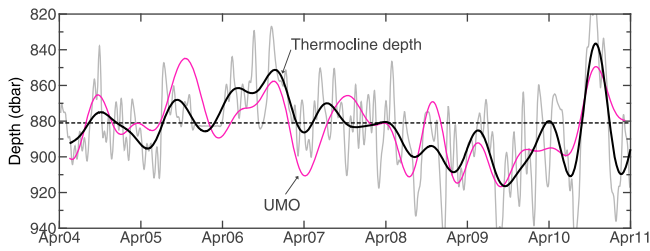
Figure 3. The average thermocline depth in decibars at the western boundary (76.75°W), here defined as the mean depth of the σ_0 surface in the range 27.4–27.45 kg m^{-3}, in grey, and smoothed with a 180-day low-pass Tukey filter in black. For reference, the low-pass filtered upper mid-ocean transport is also shown (magenta), offset and scaled.

the temperature difference is much larger leading to larger heat transports. Hence, the increasing strength of the southward upper mid-ocean transports in 09/10 marked a shift in the balance at 26°N from overturning to gyre that reduced the heat transport of the Atlantic for this period.

3.3. Changes in the Upper Mid-Ocean Transport

[15] One of the largest changes of the 09/10 period was a 2.9 Sv intensification of the upper mid-ocean transport relative to the mean from 2004–2008 (Table 1). The physical manifestation of the observed changes is a deepening

thermocline at the western boundary, particularly from mid-2005 to mid-2009. This deepening can be seen in the depth of the isopycnals in the western boundary array (see Figure 3). As the thickness of warm surface waters at the western edge of the array increased, the zonally-averaged southward return flow in the upper 1100 m intensified.

3.4. Changes in the Deep Mid-Ocean Transport

[16] The deep mid-ocean water masses at 26°N primarily consist of lower and upper NADW, of origin in the Labrador and Nordic Seas respectively. The changes in transport of these water masses during the 09/10 show major interannual variability similar to the MOC at 26°N. This shows that the variability of the NADW at 26°N is more closely related to local conditions than variability is the formation regions, which have been relatively constant over this time period [Hansen et al., 2010]. Figure 1 shows the lower NADW varying in a qualitatively similar way to the extreme Ekman transports in the latter half of 2009. However, the decline in the southward transport of lower NADW began earlier in 2009 independent of Ekman transports as shown in Figure 2 (bottom). The changes in the lower NADW transports appear primarily in the external transports (described in §2). Independent support for these changes in external transport comes from bottom pressure records. Figure 4 (top) shows that the transport anomaly derived from bottom pressures at the western boundary show high anti-correlation of $r = -0.61$, significant to 99% level, with the external transport variability, including during the reduction of the
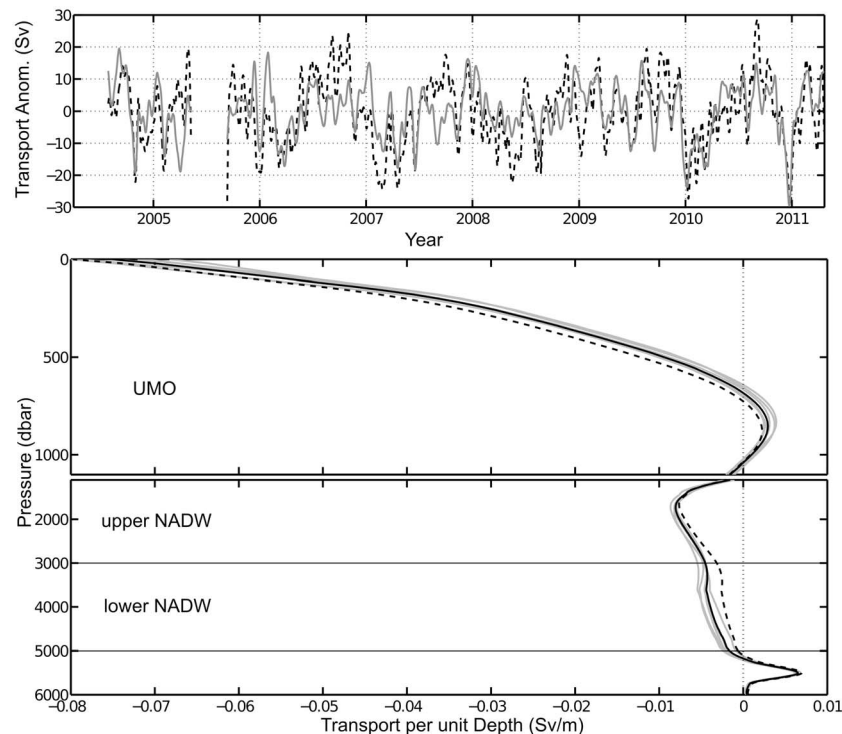
Figure 4. (top) Transport anomaly derived from the bottom pressure anomaly at 76.75°W (black, dashed) and external transport (grey). As BPR records can be subject to a linear drift, both BPR and external transport are constrained to have zero trend for each BPR deployment. Transport from the BPR assumes geostrophic balance and no compensating pressure fluctuations across the section (bottom) Transport per unit depth of upper mid-ocean profiles: 2004–2010 average (black) and 09/10 average (black, dashed). Averages for other years are in the background in grey. Note the change in vertical scale.

MOC and lower NADW transport during 09/10. This independently supports our estimate of the lower NADW transport variability, essentially extending the result of *Kanzow et al.* [2007].

[17] While the lower NADW shows the direct imprint of the downturn, the upper NADW transport was relatively unchanged during 09/10 (Table 1). A change in the shear between the upper and lower NADW layers compensated for the weakening of the lower NADW ensuring the upper NADW transport varied little. Figure 4 (bottom) shows the transport per unit depth of the mid-ocean transports, with 09/10 highlighted. The change in the upper mid-ocean transport described in §3.3 is visible in the top 1100 m. In the deep layers, the change in transport per unit depth of the lower NADW is the largest change, with a smaller change in the upper NADW from 1100–3000 m. It is not yet clear what causes this change in shear between the lower NADW and the upper NADW.

4. Conclusions

[18] During 09/10, the MOC was exceptionally low, with an average of 12.8 Sv compared to the average of 18.5 Sv from 2004 to 2008. The change is due to a combination of all the components of the circulation at 26°N but primarily in (1) the upper mid-ocean transport, which intensified with stronger southward flow from early 2009 through mid-2010 due to a deepening of the thermocline on the western boundary, and (2) the Ekman transport, which was anomalously negative from Dec 2009–Mar 2010.

[19] The changes in the upper layers were mirrored in the deep overturning. In particular, (1) the deep changes were localized in the lower NADW from 3000–5000 m, and appear as a halving of the annually averaged southward transport; (2) the upper NADW transport variability, in contrast, did not show large interannual variability due to a change in the baroclinic shear between the layers.

[20] As the MOC carries 90% of the Atlantic heat transport at this latitude [*Johns et al.*, 2011], the observations of major interannual variability in the MOC are expected to impact regional heat content in the ocean. In particular, the rebalancing of the circulation at 26°N from the deep overturning to the upper mid-ocean gyre will reduce the meridional heat transport as the ocean recirculates the warm Gulf Stream waters in the equally warm waters of the gyre rather than in the cold, deep waters of the overturning.

[21] The 26°N array continues to shed new light on the MOC at this latitude. Since 2004, these measurements of the MOC have changed perceptions of the sub-seasonal [*Cunningham et al.*, 2007], seasonal [*Kanzow et al.*, 2010; *Chidichimo et al.*, 2010] and, now, interannual variability of the MOC. We have presented evidence of a 30% reduction in the MOC which persisted for an entire year, Apr 2009–Mar 2010. While a 30% reduction certainly is large and is well outside the range predicted for interannual MOC variability in coupled ocean-atmosphere models (Figure S2 in the auxiliary material), the MOC has recovered quickly from this downturn. Further work will be needed to understand the cause of the deepening isopycnals in the thermocline and of the change in shear between the lower and upper NADW. Finally, it is clear that sustained measurements will be needed to detect any longer term decline in the MOC.

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