

# Observation of decadal change in the Atlantic meridional overturning circulation using 10 years of continuous transport data

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[1] The meridional overturning circulation (MOC) represents the main mechanism for the oceanic northward heat transport in the Atlantic, and fluctuations of this circulation are believed to have major impacts on northern hemisphere climate. While numerical ocean and climate models and paleo-records show large variability in this circulation, the use of direct observations of the MOC for detecting climate-timescale changes has proven difficult so far. This report presents the first observational record of MOC measurements that is continuous and sufficiently long to exhibit decadal-scale changes, here a decrease by 20% over the observational period (Jan. 2000–June 2009) and large interannual changes in the flow and its vertical structure. Data are from a mooring array at 16°N (Meridional Overturning Variability Experiment, MOVE). The observed change agrees with the amplitude of multi-decadal natural fluctuations seen in numerical ocean and climate models. Knowledge of the existence and phasing of such internal cycles provides multi-decadal climate predictability. Recently, some numerical model simulations have produced results that show a weakening of the MOC since the 1990's and observational confirmation of this now is a high priority. **Citation:** Send, U., M. Lankhorst, and T. Kanzow (2011), Observation of decadal change in the Atlantic meridional overturning circulation using 10 years of continuous transport data, *Geophys. Res. Lett.*, 38, L24606, doi:10.1029/2011GL049801.

## 1. Introduction

[2] The ocean circulation is a major component of the planetary heat engine, in which the Atlantic Ocean carries up to 1.3 PW of heat northward [Ganachaud and Wunsch, 2000; Johns *et al.*, 2011]. Due to its large heat storage capacity, the ocean can affect the climate of the Earth on decadal and longer timescales by creating heat imbalances due to storing, releasing, and transporting more or less heat than a steady-state would require [Ganapolski and Rahmstorf, 2001]. In the Atlantic the vast part of the heat is carried by the thermohaline or meridional overturning circulation (MOC), which is very efficient because large volumes of warm water flow northward near the surface (Gulf Stream and North Atlantic Current) and the return flow at larger depths (below about 1000 m) is much colder, by typically 10–15°C. It is expected, and demonstrated by climate models, that fluctuations in the MOC intensity cause climate variability – and due to the long timescales involved also allow climate predictability [Latif *et al.*, 2004; Keenlyside *et al.*, 2008]. The possibility of a

complete collapse of the MOC has been a subject of theoretical and modeling studies, analyses of paleo-records, and speculations [Zickfeld *et al.*, 2007].

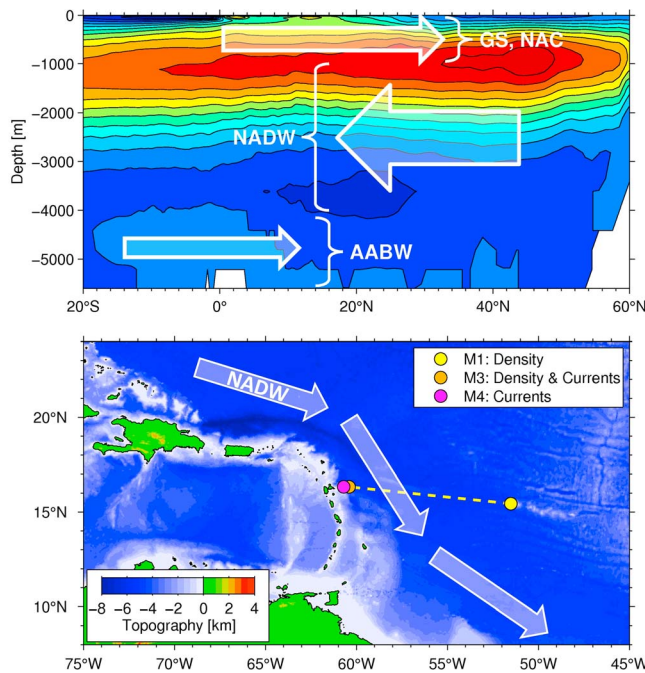
[3] In spite of the critical importance of this circulation, direct observations of its magnitude have proven very difficult [Bryden *et al.*, 2005; Wunsch and Heimbach, 2006; Willis, 2010; Cunningham *et al.*, 2007]. In principle, complete and direct flow measurements at many points across the entire Atlantic and from the surface to the bottom would be needed, and the data record needs to be accurate and long enough to detect slow changes against a background of large short-period variability. The number of flow measuring instruments required for this would be prohibitively large for operation over long times. However, meridional (i.e., north-south) transport relative to a reference level can be calculated from the difference of vertical density integrals at the east and west points of a zonal section. This “geostrophic” method has been used in oceanography for decades, but usually with ship-based CTD sections [Koltermann *et al.*, 1999], since only those allowed density computation with sufficient accuracy. Modern technology and extremely careful calibrations now make it possible to apply the geostrophic method continuously using autonomous sensors deployed on moorings for a year or more [Kanzow *et al.*, 2006, hereinafter TK06]. As a result, continuous water volume transport time-series across a section can be obtained by only instrumenting the end-points of the section with such moorings. The main unknown or assumption is the leveling of the pressure surface where the vertical integration starts between the section end points, which is equivalent to knowledge of the horizontally averaged current at that depth.

## 2. Observations

[4] We have applied this technique to the nearly 10-year long, continuous MOVE data set spanning the 1000 km section across the western basin of the Atlantic at 16°N [Kanzow *et al.*, 2008, hereinafter TK08] as shown in Figure 1 and Figure S1 in the auxiliary material.<sup>1</sup> The assertion is made that somewhere between 4300 m and 4950 m depth the long-term average meridional flow must be zero, since NADW and AABW (Figure 1) are water masses flowing in opposite directions [Bryden *et al.*, 2005]. MOVE only observes the southward NADW layer since that transport must compensate the warm, northward, upper-layer flow [Kanzow *et al.*, 2007, hereinafter TK07] (plus the much smaller northward AABW flow at the bottom). We believe this is sufficient in order to infer variability in the MOC on interannual and longer periods, also supported by numerical simulations (TK08). The main assumption we make is that virtually all of

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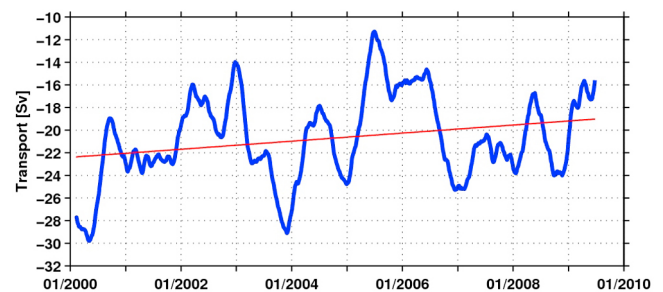
**Figure 1.** Geography and geometry of the observations. (top) Illustration of temporal mean of zonally averaged Atlantic meridional overturning stream function (i.e., mean streamlines in the N/S and vertical plane) from the JPL ECCO numerical model state estimate analysis in units of Sv ( $10^6 \text{ m}^3 \text{ s}^{-1}$ ). Contour interval is 2 Sv. Letters highlight northward flow in the Gulf Stream/North Atlantic Current system (GS, NAC) and Antarctic Bottom Water (AABW), and southward flow in the North Atlantic Deep Water (NADW). (bottom) Map showing the location of the MOVE moorings spanning the approximately 1000 km of the deep basin between coast and Mid-Atlantic Ridge, and schematic flow of NADW in the deep western boundary current (arrows). The actual flow path is more complex with recirculations and additional interior pathways.

the long-term NADW transport happens in the western basin of the Atlantic at this latitude on long timescales, which is inferred from freon (CFC) measurements [Rhein *et al.*, 2004], FLAME and GECCO model outputs for 8-year trends (see Figure S2). Results of this paper are based solely on data from three MOVE moorings (TK06): The geostrophic method is used between the end points of the 1000 km section (M1 and M3 in Figures 1 and S1) to compute across-section integrated velocities from temperature, conductivity, and pressure (and thus density) measurements. Currents on the wedge just inshore of the western mooring M3 are estimated from current meter measurements on both M3 and an additional mooring M4 on the continental slope (this configuration has been validated with prior more intensive current meter deployments). The measurement(sensor)-induced uncertainty in detecting year-to-year changes of the derived volume transport for the NADW layer is 1.3 Sv (TK08) ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ). A similar technical approach is part of the RAPID MOC mooring array at  $24^\circ\text{N}$ , which has been collecting data since 2004 [Cunningham *et al.*, 2007]. However, the RAPID-MOC covers the whole water column down to 5000 m and the whole zonal extent of the Atlantic. As a consequence, by applying mass conservation,

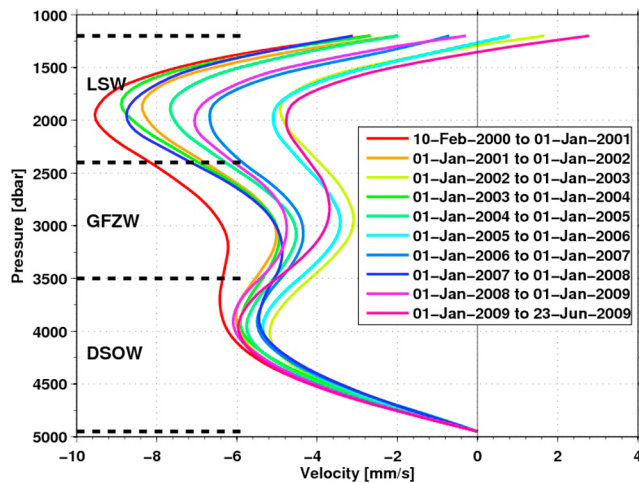
the time-variable levels of no motion near the upper and lower boundaries of the NADW are part of the transport solution in RAPID-MOC methodology [Kanzow *et al.*, 2010], whereas in the MOVE analyses the level of reference is chosen to be fixed. Thus the longer MOVE and the more encompassing RAPID-MOC observations should be considered complementary.

### 3. Results

[5] Figure 2 shows the resulting low-pass filtered time-series of NADW (1200–4950 m) water volume transport from the pressure gradients at the section end points, plus the continental slope transport from the current meters on the western moorings. It assumes that the meridional flow is zero at 4950 m (reference level, see above) based on water mass properties. The southward (negative) transport shows large interannual variability (3.8 Sv rms) and is decreasing in amplitude over the nearly 10-year record. The trend line fitted to the time series suggests a weakening by approximately 3 Sv over the entire period, which is roughly 20%. The decrease is statistically significant at 85% (see Text S1). Additional certainty comes from the vertical structure of the change (see below). Further analysis showed that the same weakening remains if a different deep reference level is chosen anywhere between 4300 and 4950 m, that it is also insensitive to the exact extent of the NADW layer which we integrate over, and to the timing (in the presence of up/down swings) of the start and end of the time-series (Figures S3 and S4). Previous studies (TK06; TK08) have demonstrated that the accuracy of the system and method is sufficient to determine the observed variability, and special attention was paid to ruling out artificial transport changes due to potential problems in instrument calibrations or due to sensor drift. The only scenario that would make the transport decrease disappear is a non-constant reference level, e.g., if the level of



**Figure 2.** 120-day low-pass filtered timeseries of the volume transport in the NADW layer (1200–4950 m) across the MOVE array (negative values denote southward flow). It is computed as the sum of the directly measured western continental slope (wedge) current and the geostrophically derived internal component of the NADW layer over the 1000 km long section, relative to 4950 m. One Sverdrup (Sv) equals  $10^6 \text{ m}^3 \text{ s}^{-1}$ . The red line is a linear fit showing a positive trend of 0.3 Sv per year, which means weakening of the southward transport. With the 120-day filter used in the figure, there are 18 degrees of freedom. Including higher frequency motions (30-day or 3-day filter) gives more degrees of freedom (32 and 40 respectively) but also more noise. In all cases a similar statistical significance of the transport decrease results of approximately 85%.



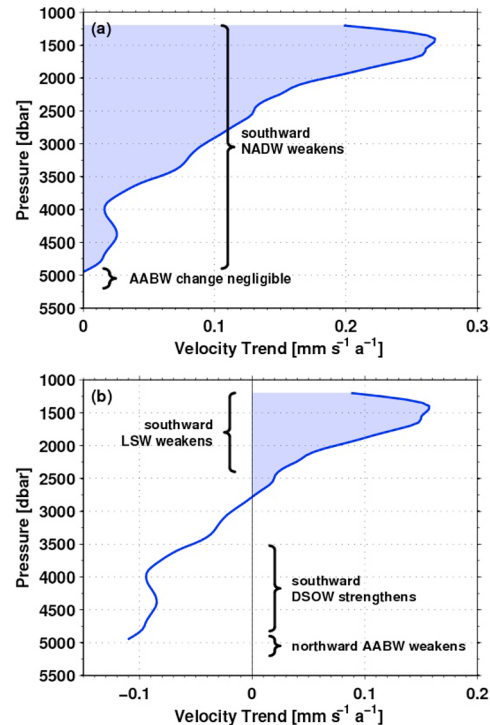
**Figure 3.** Vertical profiles of annual-mean NADW flows. Zonally averaged, geostrophically derived current across the MOVE section. By construction, the flow is always zero at 4950 dbar. Abbreviations refer to characteristic water masses and their typical depths in the MOVE data: LSW Labrador Sea Water, GFZW (Charlie-)Gibbs Fracture Zone Water, DSOW Denmark Strait Overflow Water, the water mass boundaries used are located at 1200, 2400, 3500, 4950 dbar. Colored lines show annual means over the time periods indicated, and highlight variability in the LSW layer.

zero flow were to descend by 100 m (from 4850 to 4950 m) throughout the observational period. In this scenario, transport changes are cancelled by a vertical re-distribution of the circulation, which would be an equally important change but is much less likely (see explanation in caption Figure 4b).

[6] The mooring-based time-series allows construction of vertical profiles of the horizontally averaged flow, i.e., the vertical distribution of the transport, at different times. The annual means of those profiles show large year-to-year changes in the partitioning between transports in the LSW and DSOW layers (Figure 3), with the LSW transport being weaker than the DSOW transport in 2003, 2005 and 2009, and stronger in all other years. By construction all profiles have zero speed at 4950 m depth which does not need to be satisfied on short time scales, but the vertical gradients and thus the shape of the profiles are fully determined and free of assumptions. These gradients vary much more strongly above the lower layers (above GFZW/DSOW), giving rise to large interannual changes in the flow of LSW. Analysis of the standard deviation and trend of the shear (i.e., of the vertical derivative of the transport) shows higher values above approximately 2500 m, confirming that this is the depth range with highest variability. The strengthening and weakening LSW flow agrees qualitatively with numerical ocean data assimilations [Wunsch and Heimbach, 2006].

[7] Figure 4 shows how the trend in horizontally averaged current strength is distributed over depth. The single maximum at 1400 m depth coincides with the freon-derived [Rhein et al., 2004] center of the LSW, a water mass known to exhibit large interannual and long-term changes in its formation rate [Kieke et al., 2006]. It is encouraging to see the actual transport change derived from extremely small density and pressure signals peaking in this most active NADW source layer. One might ask, whether the observed long-term

transport variability could alternatively represent a signature of a superposition of westward propagating wind-driven Rossby waves which explain much of the intraseasonal (80–250 day) NADW transport variability in MOVE (TK08). However, the vertical pattern of the change, especially the reduction of the transport change from 1400 dbar towards



**Figure 4.** Vertical distribution of transport trend, expressed as trend of the horizontally averaged current at each depth. (a) Trend in across-section geostrophic velocity as a function of depth. The integral over the shaded area comprises the positive trend shown in Figure 2. Again a positive trend implies a weakening of the southward flow in the NADW layer. The maximum near 1400 dbar corresponds very well to the known core of the LSW. Below 4950 m is the AABW (see Figure 1) which circulates northward. If the weakening of the southward NADW flow which we observe were simply a result of decreasing northward AABW (which also recirculates in the NADW layer as shown in Figure 1), then the area under the AABW curve should equal the area under the NADW layer above. The AABW extends to 5300 m in some areas on the MOVE section, thus the 300 m thick AABW is missed with the MOVE data. If this layer had a 3Sv decrease it would mean a complete halt of the AABW circulation here (it has 2–5 Sv total transport), which is oceanographically unreasonable and in contradiction with other recent observations of AABW (E. E. Frajka-Williams, personal communication, 2010). (b) Vertical profile of the trend (as in Figure 4a), if the reference zero-velocity layer were to descend by 100 m over the 10 years of observations, which would cancel the NADW transport trend. The vertical profile of the trend then would shift to the left and require the LSW flow to weaken, the DSOW transport to strengthen, and the AABW flow to weaken, just such that they all cancel. This coordinated and opposing behaviour of three independent water masses is oceanographically very difficult to rationalize and therefore deemed unlikely.



smaller depths, is difficult to reconcile with Rossby waves. The water mass, rather than Rossby wave, signature in the vertical thus gives additional significance to the transport change as a long-term, decadal signal.

#### 4. Discussion and Conclusions

[8] The main assumption made in the analysis concerns the deep reference level as a level-of-no-motion on long timescales and its constancy. On shorter (intra-annual) periods we can test the choice of reference level with absolute transport changes derived from bottom pressure measurements coupled with the density-derived transports, which reveals that the actual flow variability (without assumption of a reference level) at 4950 m is much less than at 1200 m depth (Figure S5), confirming the larger variability in the LSW layer of Figure 3 on the shorter timescales. Supporting the reference level for longer timescales, our *mean* transports agree with RAPID when using 4950 m for the lower limit of the NADW in the MOVE region. Figure 4b is used to argue (see caption) that it is unlikely that the deep level of no motion would have a descending trend of 100 m over the ten years. In addition we have analyzed the MOC intensity from various ocean models and found only small differences on 10-year timescales between using a fixed reference level or the actual time-variable level of no motion (Figure S6). Also, the observed changes in the vertical structure, the relative weakening of LSW flow, and the high variability in the LSW layer flow are not dependent on the assumption of a stationary reference level.

[9] The weakening of 3 Sv over 10 years is in order-of-magnitude agreement with a range of numerical ocean circulation models and coupled ocean-atmosphere climate models. The FLAME (TK08) and GECCO [Köhl and Stammer, 2008] models show changes of 2–4 Sv on decadal timescales, while coupled climate models like HadCM3, ECHAM5, and ARPEGE3 show fluctuations of 1.5–2 Sv over 10 years in the MOC strength [Collins *et al.*, 2006]. Thus the MOVE data would lend observational support to the magnitude of such model variability. This also suggests that the transport decrease we see in the MOC is likely to be part of natural internal multi-decadal fluctuations in the ocean-climate system rather than a secular trend, as greenhouse gas induced MOC weakening is estimated by some to amount only to 1–5% per decade [Schmittner *et al.*, 2005]. There are now some indications that ocean state estimates or reanalyses using numerical models that assimilate upper-ocean observations, exhibit decreasing MOC transport on 10-year timescales since the 1990's, at least at some latitudes, including GECCO [Pohlmann *et al.*, 2009], the SODA system (J. Carton, 2011, [www.atmos.umd.edu/~carton/](http://www.atmos.umd.edu/~carton/)), and JPL ECCO (JPL ECCO, 2011, <http://ecco.jpl.nasa.gov/external>). Transport observations with decadal variability like the ones presented here are thus urgently needed by modeling teams for comparison and validation.

[10] An important question is how representative of the large-scale pattern these MOC changes at one single latitude are. A recent study [Bingham *et al.*, 2007] has analyzed this question in two different models. While over periods of a few years or less that study finds very low coherence across latitude bands, an EOF analysis extracted a meridionally coherent larger-scale MOC mode with fluctuations at 16°N that are equivalent to 3 Sv over 10 years, in close agreement

with the 10-year variability reported here. The overall oscillations of this mode occurred on 30-year time scales. Those results imply that an MOC change such as the one observed can be meridionally coherent. Whether the change in MOC strength we observe at this latitude is indeed coherent over a range of latitudes can ultimately only be assessed in combination with other data sets. The RAPID-MOC project will also have a 10-year record four years from now, and merging the two long timeseries is important. We have just started to see some coherence between MOVE and RAPID-MOC, and this is a topic of ongoing work. Further a recent study [Lozier *et al.*, 2010] based on numerical model simulations initialized by historical hydrographic data has suggested that the MOC in subtropical and subpolar latitudes may have exhibited decadal changes of opposing signs. A coordinated network of MOC observing infrastructure covering the equatorial to subpolar Atlantic is therefore crucial.

[11] In summary, our 9.5-year time-series shows a decadal-scale change of the transport, which is robust, statistically significant with 85% confidence, gets additional certainty from its vertical distribution, is consistent with models and oceanographic knowledge, and agrees with meridionally coherent modes in models. The interannual and decadal MOC variability presented is the first direct observation of such changes in this system. Surprising year-to-year changes are also found in the vertical structure of the NADW transport. Since the weakening is concentrated in the LSW layer, it is tempting to speculate that it is related to reduced convection and water mass formation activity in the Labrador Sea since the mid 1990's, but that needs further study. If the weakening is real, this will imply a degree of predictability of northern hemisphere climate, and will allow initialization of climate forecast models with the correct MOC phase.

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