plex, and thereby protects primary synapses from destruction. This model is supported by genetic experiments showing that primary synapses are unstable in nematodes lacking SYG-1 (*syg-1* mutants) and that synapse removal requires components (SKR-1 and SEL-10) of the SCF complex.

But herein lies a paradox. In addition to protecting the primary synapse region from destruction, SYG-1 also triggers the demise of the secondary synapse region. Indeed, secondary synapses are preserved in *syg-1* mutants, whereas primary synapses are not. How does SYG-1 direct the remote destruction of the secondary synapse region? The authors show that the ubiquitin-proteasome system is required to remove secondary synapses during normal development. Thus, SYG-1 localization to the primary synapse region somehow stimulates proteolytic activity in the secondary synapse region. To explain this effect, Ding *et al*. propose that the SCF complex is limiting in the HSNL neuron. In this model, high SCF activity in the primary synapse region (for example, in mutant worms lacking SYG-1) indirectly protects the secondary synapse region from destruction. Conversely, when active SCF is excluded from the primary synapse region (as in the wild-type worms), more SCF is available to "attack" target proteins in the secondary synapse region. Although this model is pleasingly elegant and supported by additional experiments (artificial elevation of SCF activity in the HSNL neuron removes both primary and secondary synapses), Ding *et al.* acknowledge that SYG-1 could also act through unknown alternative signaling pathways to eliminate secondary synapses.

Additional specific information about the mechanism proposed by Ding *et al.* is needed to build a detailed biochemical model of the process. For example, which of the presynaptic components in the HSNL neuron is directly targeted by the SCF complex? Liprin-alpha is an attractive candidate, as its removal is predicted to destabilize presynapse assembly (*6*, *7*). Another question is whether the postsynaptic apparatus—the organization of proteins in the cells receiving stimulation by the HSNL neuron—is disassembled in concert with the presynapse. If so, how are these events coordinated?

Earlier studies have firmly established roles for the ubiquitin-proteasome system in axon guidance and synaptogenesis in disparate species. Learning and memory—higher-order functions that originate with synaptic plasticity—also depend on regulated proteolysis (*3*). Ding *et al.* have now provided an exciting example of how this degradation system can be marshaled to control the placement of specific synapses in *C. elegans*. The evolutionary conservation of the components of this mechanism suggests that additional work in this genetically tractable model organism may reveal fundamental secrets of human brain development.

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10.1126/science.1147570

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A Change in Circulation?

Measurements with a new observational array have revealed surprisingly large variations in ocean circulation in the North Atlantic.

John A. Church

Earlier this year, the Intergovernmental Panel on Climate Change (IPCC) reported unequivocal evidence that climate change is happening now (*1*). One public image of climate change is a rapid and dramatic collapse of the northward flow of warm water to high latitudes, with possible serious implications for North American and European climate. So, has the northward flow of warm water changed? How would we know if it had, and how can we monitor it in the future? A remarkable new observational program has begun to address these questions. In this issue, Kanzow *et al.* on page 938 (*2*) and Cunningham *et al.* on page 935 (*3*) report initial results from the program.

Warm water emerges from the Florida Strait and flows northward along the east coast of America as the Gulf Stream (see the figure). After leaving the coast of North America, the warm water flows northeastward before it diverges, with some water continuing northward beyond Iceland and the remainder returning in the broad southward flow of the subtropical gyre in the upper kilometer or so of the ocean.

Throughout its northward journey, the warm water loses heat to the atmosphere and becomes denser. The colder, denser water sinks at high latitudes and returns southward at depths of 2 to 5 km. This northward flow of near-surface, warm water and southward flow of cold, deep water is known as the North Atlantic meridional overturning circulation. The northward flow is controlled by a combination of surface winds and density gradients, with warmer, saltier, lighter water at low latitudes and colder, fresher, denser water at higher latitudes.

Two processes can prevent high-latitude water from sinking: surface warming and decreased salinity caused by freshwater runoff from rain and meltwater from glaciers and the Greenland Ice Sheet. This would disrupt the meridional overturning circulation. Observations indicate that there has indeed been a freshening of the North Atlantic (*4*). Climate change models (*5*) in which the CO_2 concentrations quadruple over 140 years project a decrease in the meridional overturning circulation by 10 to 50%.

Bryden *et al*. (*6*) analyzed five sets of ship-based, full-depth temperature and salinity measurements across the North Atlantic near 25°N, completed between 1957 and 2004. From these data, they estimated that the overturning circulation had decreased by 30% and the northward heat transport decreased by over 20%, as a result of an inferred decrease in the southward transport of the coldest, densest water and an increase in southward transport of the warmer water in the upper kilometer of the ocean. This decrease in the meridional overturning circulation is much larger than suggested by climate model simulations of the 20th century, and more akin to the up to 50% decrease projected for the end of the 21st century (*5*, *7*). Are the models not responding rapidly enough to greenhouse gas changes, or was the observed change—as Bryden *et al.* cautioned—uncomfortably close to the uncertainties in previous observational estimates?

In a bold new initiative led by the UK

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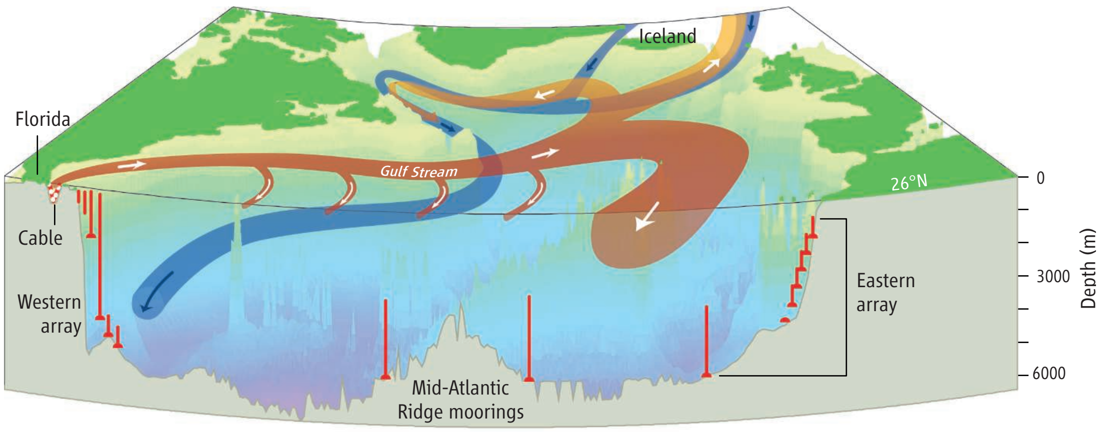

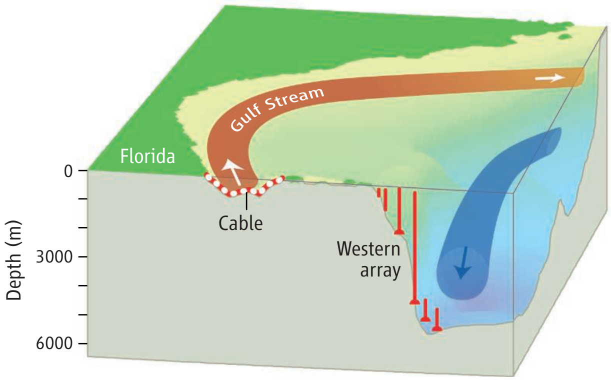

Effective monitoring. The RAPID/MOCHA monitoring array, operational since 2004, directly measures the transport of the Gulf Stream in the Florida Strait using an undersea cable (red). A sparse moored array along 26°N measures bottom pressure and water column density at the western and eastern boundary and on either side of the Mid-Atlantic Ridge; in the lower panel, the array is shown in detail for the western boundary. Two articles in this issue (*2*, *3*) report results for the first year of the array's operation.

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National Environmental Research Council (with support from U.S. agencies), the RAPID/MOCHA (Rapid Climate Change/Meridional Overturning Circulation and Heat Flux Array) array was deployed in March 2004 to continuously monitor the meridional overturning circulation at 26°N. These observations are a component of a larger set of activities of the World Climate Research Programme's CLIVAR (Climate Variability and Predictability) Project to monitor the North Atlantic meridional overturning circulation.

The inexpensive array along 26°N consists of instruments measuring the variations in the bottom pressure, and temperature and salinity (and thus density), throughout the water column near the western and eastern boundaries and on either side of the Mid-Atlantic Ridge (see the figure). These measurements can be combined to estimate variations in the horizontal pressure difference between the western and eastern boundary throughout the water column. The pressure differences are directly proportional to variations in the horizontally integrated northward flow. The ocean-interior measurements are complemented by estimates of the northward flow of the Gulf Stream through the Florida Strait by an undersea cable and satellite measurements of the wind stress and hence of the surface-wind–driven transport.

On time scales of 15 days and longer, the sum of transports into the North Atlantic should be about zero. Indeed, the observations reported by Kanzow *et al.* indicate that the sum varies with a root-mean-square value of only 3.4 Sv (1 Sv = 10^6 m^3/s), slightly larger than the expected measurement errors of 2.7 Sv, thus demonstrating the remarkable effectiveness of the array. The fact that the observed sum varies slightly more than the expected measurement errors presumably reflects deficiencies in the method, such as the unobserved flow deeper than the deepest part of the array and the impact of the Mid-Atlantic Ridge.

Cunningham *et al.* report a year-long average meridional overturning circulation of 18.7 ± 5.6 Sv (*3*), but with large variability ranging from 4.4 to 35.3 Sv over the course of the year. This range includes all five meridional overturning circulation values estimated from the snapshots analyzed by Bryden *et al.*; thus, the apparent long-term decrease inferred by these authors may merely be a result of large intra-annual variability.

Cunningham *et al.* estimate that they can measure the annual average overturning to a resolution of 1.5 Sv. This would be sufficient to detect any large abrupt transition of the meridional overturning circulation. An assessment of the current generation of climate models indicates that such a large abrupt transition is very unlikely during the 21st century (*6*), but verification of these projections and continuing assessment of the stability of the meridional overturning circulation remains a priority.

It remains unclear how much the meridional overturning circulation varies from year to year. Understanding this variability will be critical to improving models, thus allowing more reliable projections of climate change. This variability will determine how long a record will be required to determine a trend in the meridional overturning circulation. A recent coupled ocean atmosphere model study (*8*) suggests that it would take several decades of observations to detect such a trend. Similarly, it will take decades of monitoring to determine which (if any) of the models analyzed by the IPCC (*7*) most accurately reflects reality.

The effectiveness and the inexpensive nature of the RAPID/MOCHA array should allow long-term monitoring of an important element of the global climate system. Equivalent observational schemes for the Southern Ocean limb of the meridional overturning circulation, where decadal water-mass changes have also been observed (*9*), remain to be designed.

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10.1126/science.1147796

Downloaded from www.sciencemag.org on October 7, 2010

Published by AAAS

Temporal Variability of the Atlantic Meridional Overturning Circulation at 26.5°N

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The vigor of Atlantic meridional overturning circulation (MOC) is thought to be vulnerable to global warming, but its short-term temporal variability is unknown so changes inferred from sparse observations on the decadal time scale of recent climate change are uncertain. We combine continuous measurements of the MOC (beginning in 2004) using the purposefully designed transatlantic Rapid Climate Change array of moored instruments deployed along 26.5°N, with time series of Gulf Stream transport and surface-layer Ekman transport to quantify its intra-annual variability. The year-long average overturning is 18.7 ± 5.6 sverdrups (Sv) (range: 4.0 to 34.9 Sv, where 1 Sv = a flow of ocean water of 10^6 cubic meters per second). Interannual changes in the overturning can be monitored with a resolution of 1.5 Sv.

Defining the size of the variability in MOC is a fundamental prerequisite to understanding how it may be changing on climate-relevant time scales. Recently, it was suggested that the circulation has slowed by 30% or 8 Sv over the past decade, based on consistent analysis of repeated hydrographic sections along 26.5°N (*1*). Coupled climate models suggest that there will be a decline in the Atlantic overturning circulation as a result of increased CO_2 in the atmosphere but that the change will be gradual over this century (*2*). Is the suggested 8-Sv change in overturning just the result of intra-annual variability in the circulation sampled by each hydrographic section? What size of a change in the overturning can be reliably detected above the intra-annual variability? To answer such questions, we define the size and structure of the intra-annual variability in the MOC at 26.5°N from 1 year of measurements by using the Rapid Climate Change (RAPID) mooring array (*3*).

The 26.5°N Atlantic section is separated into two regions: a western boundary region, where the Gulf Stream flows through the narrow (80 km), shallow (800 m) Florida Straits between Florida and the Bahamas, and a transatlantic mid-ocean region, extending from the Bahamas at about 77°W to Africa at about 15°W (fig. S1). Variability in Gulf Stream flow is derived from cable voltage measurements across the Florida Straits (*4*), and variability in wind-driven surface-layer Ekman transport across 26.5°N is derived from QuikScat satellite–based observations (*5*). To monitor the mid-ocean flow, we deployed an array of moored instruments along the 26.5°N section (fig. S2). The basic principle of the array is to estimate the zonally integrated geostrophic profile of northward velocity on a daily basis from time-series measurements of temperature and salinity throughout the water column at the eastern and western boundaries. Inshore of the most westerly measurements of temperature and salinity, the transports of the Antilles current and deep western boundary current are monitored by direct velocity measurements [supporting online material (SOM) text].

We deployed the mid-ocean array from February to March 2004 and recovered it from March to May 2005 (*6*, *7*). The overlapping period when the entire array was working for its first year was 28 March 2004 to 31 March 2005. We have since redeployed the array in spring 2005 and again in spring 2006. In this report, we present results from the first year. The design of the RAPID array for monitoring basin-scale circulation was tested in numerical ocean-circulation models (*8*, *9*), and a companion paper (*10*) demonstrates from the first year's time series that five independently measured transports (Gulf Stream, Ekman, boundary wedge, baroclinic, and barotropic geostrophic transports) are in overall mass balance for time scales longer than 10 days, providing evidence that the monitoring system works (SOM text).

To examine the intra-annual mid-ocean baroclinic variability in layers (Fig. 1), we estimated daily transports above 800-m depth (thermocline recirculation), between 800- and 1100-m depth (intermediate water flow), between 1100- and 3000-m depth (upper North Atlantic deep water or UNADW), and below 3000-m depth (lower North Atlantic deep water or LNADW). We prefer to use 800 m as a boundary for the thermocline recirculation because this is the maximum depth of the Florida Straits; thus, all of the northward transport in the Gulf Stream and Ekman layer occurs above 800-m depth.

The time series of layer transports (Fig. 2) exhibit variability of about 3 Sv around their time-averaged transports (Table 1): The SD in thermocline recirculation and LNADW is ±2.7 and ±3.5 Sv, respectively, indicating the size of the variability in baroclinic structure. Such variability is smaller than the 6 Sv previously reported from a modeling study (*11*). Still, the range in transports is large: The southward thermocline recirculation is as small as –6.6 Sv and as large as –23.3 Sv, and the range in LNADW transport is from 1.0 to –18.2 Sv.

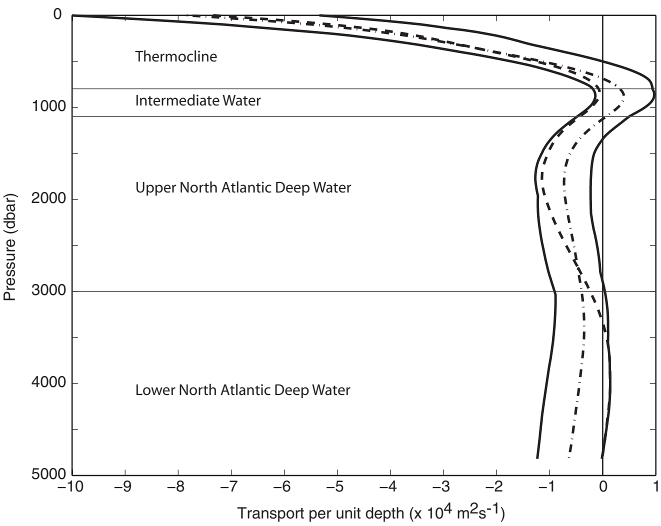

Fig. 1. Vertical profile of the northward mid-ocean transport per unit depth (m^2 s^{-1}). Dynamic height difference (east minus west in m^2 s^{-2}) divided by Coriolis frequency (s^{-1}) equals transport per unit depth and is proportional to the zonally integrated meridional geostrophic velocity; negative difference corresponds to southward velocity across 26.5°N. Reference levels are chosen so that the vertically integrated mid-ocean geostrophic transport equals the northward Gulf Stream transport through the Florida Straits plus the average northward wind-driven Ekman transport across 26.5°N plus the boundary wedge transport. The profile of year-long average (dashed-dotted curve), the profile on 2 November 2004 during the extreme (dashed curve), and the maximum and minimum over the year at each depth (solid curves) are shown.

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From March 2004 to March 2005, the Florida Straits transport is at its maximum in August (Fig. 3), typical of its long-term seasonal cycle (*12*). The year-long average transport is 31.7 Sv, slightly less than the long-term mean of 32.2 Sv (*12*), and the daily transport variations have a SD of ±3.3 Sv (Table 2). During 2004–2005, the northward wind-driven surface-layer Ekman transport has an annual average value of 3.0 Sv, smaller than the long-term mean Ekman transport of about 3.8 Sv, due in part to anomalous southward transport during February and March 2005. Long-term records of Ekman transport at 26.5°N exhibit small seasonal variability (*13*), but from 2004 to 2005, the maximum northward Ekman transport occurs in December and January. Daily variations in Ekman transport via QuikScat data have a SD of ±4.4 Sv. The mid-ocean thermocline recirculation above 800-m depth is at a minimum in late September and a maximum in December.

The maximum overturning, as is commonly used by modelers in their overturning analyses, is defined here as the sum of northward Gulf Stream, Ekman, intermediate water, and southward thermocline recirculation transports, which gives the maximum amount of northward transport of upper waters (SOM text); maximum overturning has an annual mean transport of 18.7 Sv with a SD of ±5.6 Sv. The overturning reaches a maximum value of 34.9 Sv in September, when the Gulf Stream transport is near its summertime maximum and when the southward upper mid-ocean transport is near its minimum value. The overturning achieves a minimum value of 4.0 Sv in February when the Gulf Stream transport is low, the Ekman transport is southward, and the southward upper mid-ocean transport is relatively strong.

Gulf Stream, Ekman, and upper mid-ocean transport (Fig. 3) are nearly independent time series, and there is no significant correlation among them. There is some compensation inherent in the mid-ocean recirculation because the reference-level velocity depends on the size of the Gulf Stream transport plus Ekman transport on each day, but most of the transport compensation occurs in the deep-water transports that have larger areas. The SD in upper mid-ocean transport (±3.1 Sv) is actually smaller than the variations in Gulf Stream (±3.3 Sv) transport or Ekman (±4.4 Sv) transport. Therefore, the variance in the overturning is nearly equal to the sum of the variances, so each component contributes about equally to its temporal variability.

The most notable event in the year-long time series occurs in early November 2004, when the deep southward flow of LNADW essentially ceased and there was a brief period of net northward transport of deep waters below 3000-m depth. In the time series of temperatures at the western boundary station (fig. S4), the signature of this event is a 700-m downward displacement of isotherms below 2000 m. The vertical profile of dynamic height difference (Fig. 1) shows the sharp decrease in southward velocity in the deep water, as compared to the average profile. In terms of water masses, the cold LNADW effectively disappears at the boundary. Even though the southward flow of LNADW stopped, there is not a large anomaly in overturning. This event is strongly baroclinic (Figs. 1 and 2): The thermocline recirculation is close to its average value whereas the southward flow of UNADW is larger than average, which compensates for the lack of LNADW transport. Thus, the overturning (Fig. 3) is close to its mean value during this event.

There has been no comparable event observed in historical transatlantic sections (*14*–*18*). There was a steep descent of deep isotherms offshore from the western boundary in the 1957 hydrographic section, but the isotherms recovered

Fig. 2. Year-long time series of layer transports for thermocline recirculation (red), intermediate water (green), UNADW (light blue), and LNADW (dark blue). Negative transports correspond to southward flow.

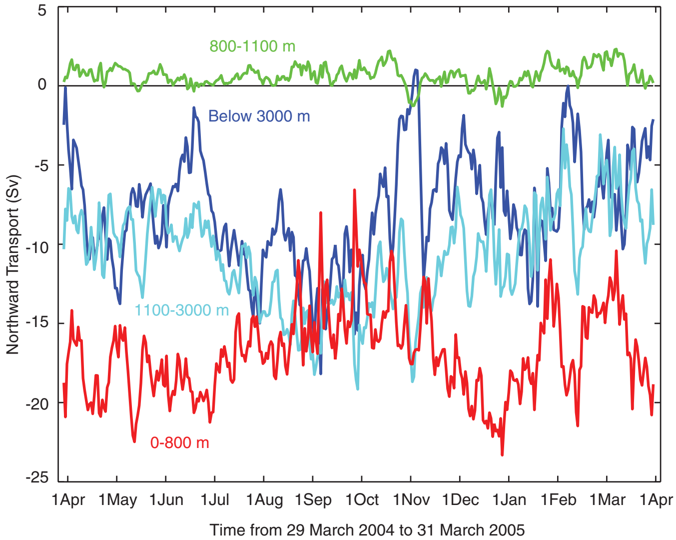
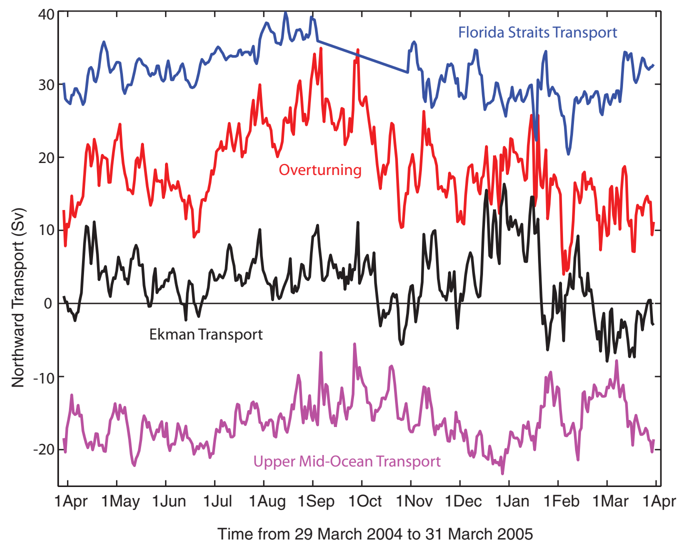


Table 1. Mid-ocean layer transports (data reported in Sv).

Water depth	Mean	SD	Minimum	Maximum
0 to 800 m (thermocline recirculation)	−16.9	2.7	−23.3	−6.6
800 to 1100 m (intermediate water)	0.7	0.6	−1.3	2.3
1100 to 3000 m (UNADW)	−10.7	3.1	−19.2	−2.7
Below 3000 m (LNADW)	−7.8	3.5	−18.2	1.0

Fig. 3. Daily time series of Florida Straits transport (blue), Ekman transport (black), upper mid-ocean transport (magenta), and overturning transport (red) for the period 29 March 2004 to 31 March 2005. Florida Straits transport is based on electromagnetic cable measurements; a gap in the time series of ~2 months from 4 September to 28 October 2004 is due to Hurricane Frances, which destroyed the facility recording the voltage (linear interpolation is chosen here to fill the gap). Ekman transport is based on QuikScat-determined winds. The upper mid-ocean transport, based on the RAPID time series, is the vertical integral of the transport per unit depth down to the deepest northward velocity (~1100 m) on each day. Overturning transport is then the sum of the Florida Straits, Ekman, and upper mid-ocean transports and represents the maximum northward transport of upper-layer waters on each day (SOM text).



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at the boundary, suggesting that there was a deep eddy of recirculating LNADW in this offshore region. With the western station of the RAPID array being so close to the western boundary, there is no room for a deep western boundary current to be inshore of mooring WB2 (SOM text), so we do not consider this November 2004 event to be an eddy. Within 500 km of the Bahamas, the deep western boundary current transport is absent at this time.

In a recent paper, the strength of the MOC calculated from hydrographic sections in 1957, 1981, 1992, 1998, and 2004 was found to have reduced from 22.9 to 14.8 Sv, where the overturning was defined to be the net northward transport above 1000-m depth (*1*). For the 2004–2005 time series, the overturning shallower than 1000 m has a mean value of 19.0 Sv and a SD of ±5.6 Sv. For the year-long time series, the range in daily overturning includes all hydrographic section estimates, suggesting that single hydrographic sections may represent only intra-annual variability rather than a long-term trend. The 2004 hydrographic section started at the western boundary on 7 April 2004, when the thermocline recirculation was large, when LNADW transport was small (Fig. 2), and when the overturning was small (Fig. 3). Thus, relative to the 2004–2005 time series, the 2004 hydrographic section was taken during a period of low overturning relative to the year-long average overturning.

The temporal variability in the overturning resulting from fluctuations in the Florida Straits or Ekman transports can be put into the context of long time series of such transports. The Florida Straits transport time series goes back reliably to 1982, with some cable time series continuing to the 1950s and some dropsonde sections extending to as early as 1964 (*12*, *19*, *20*). Similarly, the National Centers for Environmental Prediction reanalysis project (*21*) archives wind stress values back to the 1950s. Thus, any variability in annually averaged overturning, due to changes in Florida Straits or Ekman transports of more than 1 or 2 Sv, should be immediately recognizable. However, we lack a comparable long time series of mid-ocean transport against which we might examine changes in mid-ocean circulation that would affect the strength of the overturning. There are isolated hydrographic stations on the eastern and western boundaries that provide some additional information on long-term variability in mid-ocean transport (*22*), but these "snapshot" estimates of mid-ocean circulation all lie within the range of variability in the first year's RAPID measurements. Without additional historical estimates to increase the degrees of freedom, it is unlikely that we will be able to conclusively demonstrate a change in the overturning circulation over the past 50 years.

In terms of the future detection of changes in overturning, the prospects are better. The 2004–2005 time series define a year-long average upper mid-ocean transport of –16.1 Sv with a SD of ±3.1 Sv. Based on the integral time scales of variability (SOM text), the SE of the yearly average mid-ocean transport is about 0.8 Sv. Indeed, we can monitor the yearly average mid-ocean layer transports as well as the Florida Straits or Ekman transports. If future years' time series exhibit similar intra-annual variability, we should be able to identify real interannual variability that is larger than 1.6 Sv in mid-ocean transport averaged over a year. Combining the mid-ocean Ekman and Florida Straits transports into a time series of the overturning, we can estimate the yearly average overturning with a SE of about 1.5 Sv. Thus, we can monitor the interannual variability in the overturning at 26.5°N with a resolution of 1.5 Sv. For example, if the circulation passes through a bifurcation (*23*) or if the overturning reduces by 25% [as coupled climate models suggest that it might under increasing atmospheric CO_2 concentrations (*24*)], we should also be capable of identifying the change relative to the 2004–2005 average. There may be interannual variability in the circulation and overturning that would obscure a trend, as found in coupled climate models (*25*); but, with longer time series of mid-ocean transports from the RAPID array (combined with continuing cable measurements of Florida Straits transport and satellite-based wind estimates of Ekman transport), the interannual variability in Atlantic overturning should be defined with a resolution of 1.5 Sv.

Thus, although the intra-annual variability in the overturning demonstrates that we are unlikely to conclusively identify past changes in the overturning using only sparse basin margin densities at a single latitude, the observed temporal variability defines the limit of our ability to identify future changes. Fundamentally, we need longer time series of the overturning to define its interannual variability. Ten additional years of uninterrupted measurements would ensure that any seasonal cycles are well defined and would also refine the nature of interannual variations, whether they are oscillations, trends, or sudden shifts.

Table 2. Component transports of the Atlantic overturning circulation (data reported in Sv) at 26.5°N for the period 29 March 2004 to 31 March 2005. Overturning transport is defined as in Fig. 3. The upper mid-ocean transport is defined as the minimum in southward transport of upper waters on each day. The average depth of the maximum transport is 1041 m (with a SD of ±92 m).

Component	Mean	SD	Minimum	Maximum
Florida Straits transport	31.7	3.3	20.4	39.8
Ekman transport	3.0	4.4	–7.9	16.3
Upper mid-ocean transport	–16.1	3.1	–23.3	–5.5
Overturning	18.7	5.6	4.0	34.9

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26. The authors would like to thank the captains and crews of the research vessels *Charles Darwin*, *Discovery*, *Ronald Brown*, and *Knorr*, and the UK National Marine Facilities-

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Sea Systems mooring team. The mooring operations are funded by NERC and NSF. The Florida Current cable data are made freely available by the AOML (www.aoml.noaa.gov/phod/floridacurrent/) and are funded by the NOAA Office of Climate Observations. Wind stress data were obtained from the Centre for Satellite Exploitation and Research, at the Institut français de recherche pour l'exploitation de la mer, Plouzané, France.

Supporting Online Material
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14 February 2007; accepted 12 July 2007
10.1126/science.1141304

Observed Flow Compensation Associated with the MOC at 26.5°N in the Atlantic

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The Atlantic meridional overturning circulation (MOC), which provides one-quarter of the global meridional heat transport, is composed of a number of separate flow components. How changes in the strength of each of those components may affect that of the others has been unclear because of a lack of adequate data. We continuously observed the MOC at 26.5°N for 1 year using end-point measurements of density, bottom pressure, and ocean currents; cable measurements across the Straits of Florida; and wind stress. The different transport components largely compensate for each other, thus confirming the validity of our monitoring approach. The MOC varied over the period of observation by ±5.7 × 10^6 cubic meters per second, with density-inferred and wind-driven transports contributing equally to it. We find evidence for depth-independent compensation for the wind-driven surface flow.

The Atlantic meridional overturning circulation (MOC) consists of a near-surface, warm northward flow, compensated for by a southward return flow at depth. Heat loss to the atmosphere makes the increasingly dense northward-flowing surface waters sink at high latitudes to feed the deep return flow (*1*). The vertical temperature contrast associated with this flow results in a northward heat transport of 1.3 × 10^{15} W at 24°N (*2*), which noticeably moderates the Northeast Atlantic climate (*3*, *4*).

Most of the observation-based estimates of Atlantic MOC strength are based on infrequently acquired zonal hydrographic sections. Because the frequency distribution of the MOC variability is unknown, long-term changes inferred from these snapshot sections (*5*) may not be representative. Basic MOC characteristics, such as magnitude and time scales of natural variability (*6*), response to local wind-stress forcing, or the relative importance of wind-stress and buoyancy forcing on subseasonal-to-decadal time scales (*7*, *8*), have not yet been observed. Our ability to detect future MOC changes depends on the accurate quantification of the MOC's spectral distribution and on understanding the physical processes involved.

We analyzed MOC variability on subseasonal time scales using a 1-year-long mooring-based volume-transport time series from March 2004 to March 2005, acquired in the framework of the rapid climate change/meridional overturning circulation and heat flux array (RAPID/MOCHA) experiment (*9*, *10*). To compute the MOC, the zonally integrated meridional flow across 26.5°N as a function of depth (z) was observed. The backbones of this effort are moorings that measure full water-column profiles of density and ocean-bottom pressure at the western and eastern endpoints of the basin interior (Fig. 1) and on both sides of the Mid-Atlantic Ridge (MAR) (fig. S2). The eastern-to-western boundary-density difference allows for the computation of the temporal evolution of the basin-wide integrated geostrophic-transport profile relative to 4820 dbar

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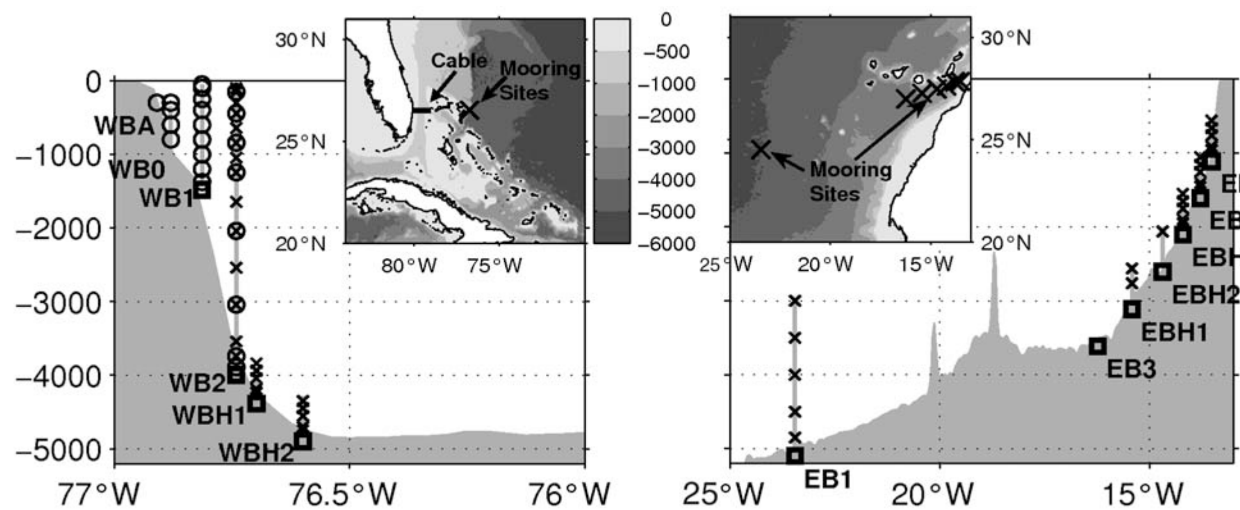

Fig. 1. Distribution of density (crosses) and bottom-pressure sensors (squares) of the RAPID/MOCHA moorings at the western and eastern boundaries of the subtropical North Atlantic near 26.5°N that are used for computing the zonally integrated meridional geostrophic flow. Direct current–meter measurements at the western boundary (circles) complement the observations in the upper part of the western-boundary continental slope. The location of the western- and eastern-boundary mooring sites and that of the Straits of Florida telephone cable can be seen in the insets. WBA, western boundary acoustic doppler current profiler; WBH, western boundary homer; EB, eastern boundary.