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Reduced Antarctic meridional overturning circulation reaches the North Atlantic Ocean

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[1] We analyze abyssal temperature data in the western North Atlantic Ocean from the 1980s–2000s, showing that reductions in Antarctic Bottom Water (AABW) signatures have reached even that basin. Trans-basin oceanographic sections occupied along 52°W from 1983–2003 and 66°W from 1985–2003 quantify abyssal warming resulting from deepening of the strong thermal boundary between AABW and North Atlantic Deep Water (NADW), hence a local AABW volume reduction. Repeat section data taken from 1981–2004 along 24°N also show a reduced zonal gradient in abyssal temperatures, consistent with decreased northward transport of AABW. The reduction in the Antarctic limb of the MOC within the North Atlantic highlights the global reach of climate variability originating around Antarctica. **Citation:** Johnson, G. C., S. G. Purkey, and J. M. Toole (2008), Reduced Antarctic meridional overturning circulation reaches the North Atlantic Ocean, *Geophys. Res. Lett.*, *35*, L22601, doi:10.1029/2008GL035619.

1. Introduction

[2] The deep ocean is important to Earth's climate, storing substantial anthropogenic heat [*Levitus et al.*, 2005], contributing to sea level rise [*Domingues et al.*, 2008], and globally transporting heat, freshwater, and biogeochemical parameters. Shutdown of the North Atlantic Deep Water (NADW) sinking limb of the Meridional Overturning Circulation (MOC) is a proposed initiator of past rapid climate change [*Broecker*, 1998]. However, the other MOC limb, fed by sinking Antarctic Bottom Water (AABW), is of comparable size [*Orsi et al.*, 1999] with AABW covering about twice as much ocean floor as NADW and occupying about twice its volume [*Johnson*, 2008].

[3] As it spreads northward into each of the three major oceans, AABW mixes with overlying waters, but nevertheless, AABW influences can be traced even to the western basins of the North Atlantic [*Orsi et al.*, 1999; *Johnson*, 2008]. AABW has warmed around the globe over the past few decades. Weddell Sea Bottom Water, the coldest, densest variety of AABW, warmed and lost volume during the 1990s [*Fahrbach et al.*, 2004]. From the 1970s through the 1990s Warm Deep Water, another AABW constituent, also warmed by $\sim 0.01°C\ yr^{-1}$ in the Ross [*Jacobs et al.*, 2002] and Weddell [*Robertson et al.*, 2002] seas. Downstream of the AABW formation regions, comparisons of potential temperature (θ) data collected on sections repeated one or more times since the 1980s reveal abyssal warming in deep basins ventilated by AABW in the Southeast Indian [*Johnson et al.*, 2008], Pacific [*Fukasawa et al.*, 2004; *Kawano et al.*, 2006; *Johnson et al.*, 2007], western South Atlantic [*Coles et al.*, 1996; *Johnson and Doney*, 2006; *Zenk and Morozov*, 2007], and even equatorial Atlantic [*Hall et al.*, 1997; *Andrié et al.*, 2003] oceans, although in the last not unambiguously [*Limeburner et al.*, 2005]. In addition, the strong vertical θ gradient (the deep thermocline) between NADW and AABW deepened in the South Atlantic and on the equator [*Johnson and Doney*, 2006; *Limeburner et al.*, 2005]. A reduction of AABW influence, communicated not by advection but by Kelvin and Rossby Waves in the presence of local lateral and vertical gradients, could produce such warming on these timescales [*Nakano and Suginohara*, 2002].

[4] Here ten data sets from three sections repeatedly occupied in the western North Atlantic (Figure 1) are analyzed with respect to AABW changes. Two repeated meridional sections [*Joyce et al.*, 1999], one along 52°W (occupied in 1981, 1997, and 2003) and the other along 66°W (occupied in 1985, 1997, and 2003), both reveal AABW warming. In addition, a repeated zonal section [*Bryden et al.*, 2005; *Longworth*, 2007] along 24°N (occupied in 1981, 1992, 1998, and 2004) shows a reduction in northward AABW transport.

2. Data and Analysis Methods

[5] For this study modern post-1980 oceanographic data from repeat sections along 52°W, 66°W, and 24°N are analyzed to assess time variability of abyssal ocean θ, its statistical significance, and geostrophic volume transport variability. The 1980s sections were occupied during the ramp-up to the World Ocean Circulation Experiment (WOCE). The 1990s sections were occupied during WOCE. The 2000s sections were occupied in support of CLIVAR (Climate Variability) and Carbon Cycle Science Programs.

[6] Vertical profiles of specific volume anomaly and θ on the 1968 International Practical Temperature Scale are calculated from the data at each station. The quantities are then interpolated onto a closely spaced pressure-latitude (or pressure-longitude) grid and masked using bottom bathymetry before differences, means, or velocities are calculated. Degrees of freedom for the differences are estimated from integrals of their spatially lagged autocorrelations [*von Storch and Zwiers*, 2001] at each pressure. Student's T-test is then applied to estimate the 95% confidence intervals.

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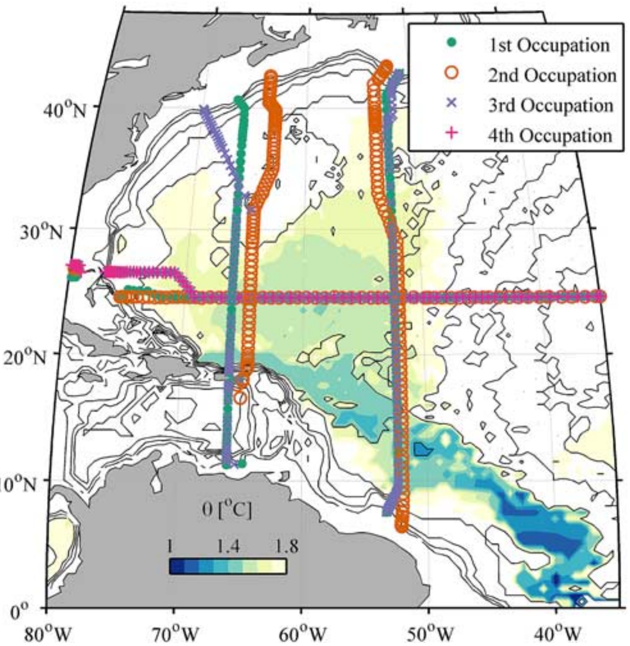

Figure 1. Station locations from repeat oceanographic sections along 24°N, 52°W, and 66°W (see legend) plotted over climatological bottom potential temperature [*Gouretski and Koltermann*, 2004], color shaded where θ < 1.8°C. Bathymetry at 1000-m intervals (thin lines) and land (grey shading) are also indicated.

[7] The data from these sections are especially useful for detecting subtle changes in abyssal temperatures. These sections have relatively close station spacing, with nominal horizontal distances generally 55 km between stations. They also all employ a CTD (Conductivity-Temperature-Depth) instrument, affording very accurate (±0.002°C) vertically continuous temperature profiles associated with very accurate (±2 dbar) pressure measurements from the sea surface to the ocean floor [*Joyce et al.*, 1999].

[8] To ensure comparison of only geographically co-located data, at each longitude (or latitude) of the grid, only data shallower than the shallowest of a set of straight lines (one for each section) connecting the deepest sample pressures of adjacent stations versus longitude (or latitude) are used in the analysis. In addition to this mask, data at grid points exceeding the pressure corresponding to a bathymetric estimate [*Smith and Sandwell*, 1997] for each location are also discarded. Small deviations in individual section longitudes from 52°W and 66°W and latitudes from 24°N (Figure 1) are ignored. However, calculations of θ differences and geostrophic velocities are generally limited to regions where deviations from the nominal section longitude (or latitude) are small.

3. Results

[9] Consistent with climatological bottom θ [*Gouretski and Koltermann*, 2004] in the western North Atlantic (Figure 1), the strongest AABW influence along the 52°W section is at its southern end, from 8–18°N, and below 4500 dbar (Figure 2a). There the strong abyssal thermocline

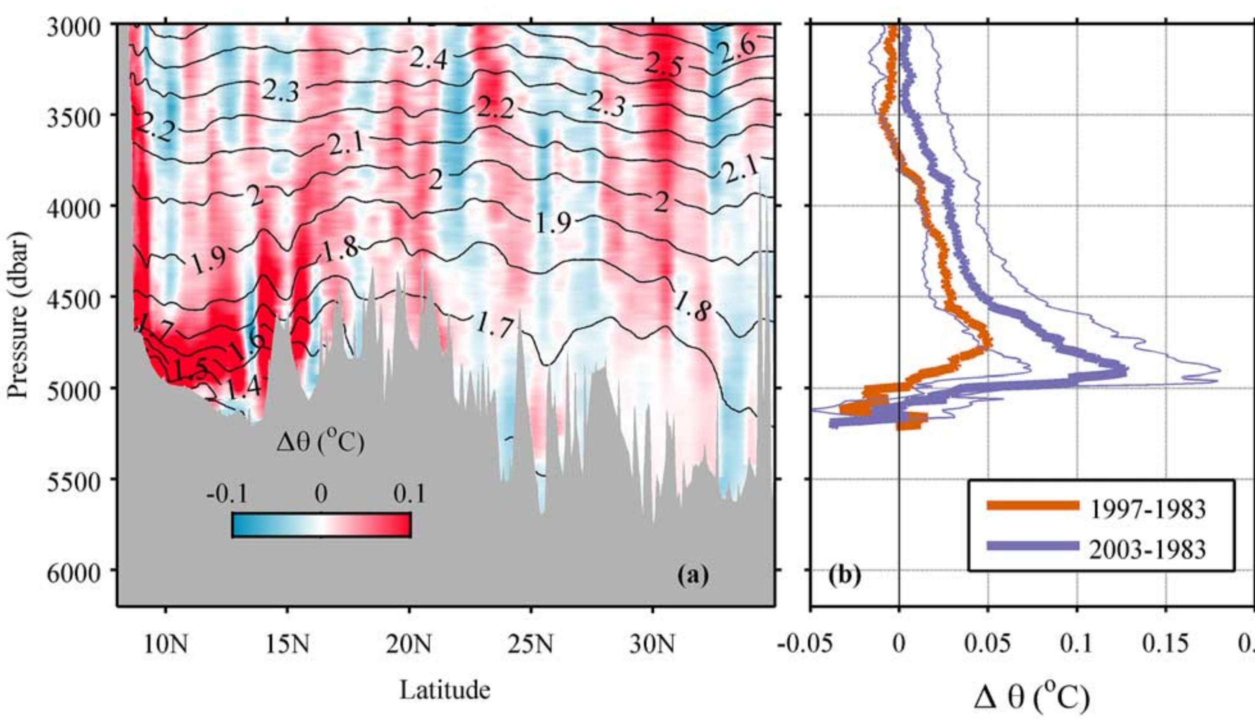

Figure 2. Deep θ variability along 52°W (Figure 1). (a) Differences in θ for 2003–1983 (color shading) south of 35°N contoured versus pressure and latitude with mean potential isotherms (black contours) from 1983, 1997, and 2003 data overlaid and bottom bathymetry (grey shading) indicated. (b) Mean θ differences from 8–18°N, for 1997–1983 (orange line) and 2003–1983 (thick blue line) with 95% confidence interval (thin blue lines).

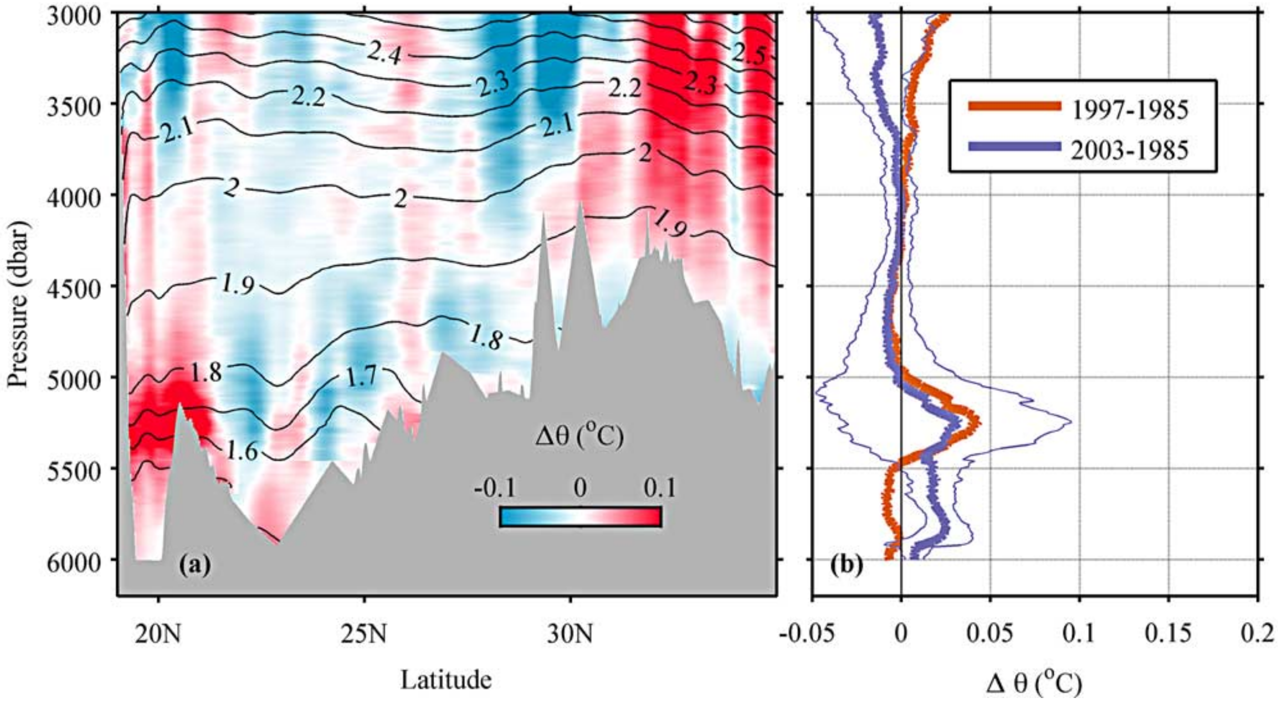

Figure 3. Deep θ variability along 66°W (Figure 1). (a) Following Figure 2a but for 2003–1985 along 66°W from 18–35°N with mean potential isotherms from the 1985, 1997, and 2005 data. (b) Following Figure 2b, but from 18–27°N along 66°W for 1997–1985 and 2003–1985.

from 4500–5000 dbar marks the vertical transition between NADW and AABW. Only from 8–18°N do the time-mean near-bottom θs drop below 1.5°C, indicating strong AABW influence. To the north, where the AABW influence is diminished, near-bottom θs are much higher.

[10] Subtracting gridded 1983 θs from 2003 values at each location (Figure 2a) quantifies warming exceeding 0.1°C on average from 8–18°N near 4900 dbar (Figure 2b). The abyssal warming, significantly different from zero at 95% confidence from 3800–5000 dbar, is caused by 80–150 dbar sinking of the abyssal thermocline in this region. The deep 1997–1983 section-mean differences from 8–18°N lie between the 2003–1983 values, consistent with monotonic warming.

[11] Along 66°W, AABW influence is evident from about 18–27°N (Figure 3a), where bottom θs are coldest, and the abyssal thermocline strongest. However, everywhere along 66°W the time-mean near-bottom θs exceed 1.4°C, warmer than values seen along 52°W, as in climatological bottom θ (Figure 1). The abyssal thermocline at 66°W, located from 5000–5550 dbar, is weaker than at 52°W, and is also deeper by ~500 dbar. Unlike at 52°W, a relatively homogenous bottom layer in θ, ~500-dbar thick, fills the southern end of the basin at 66°W. These features are signatures of an AABW layer cascading downward and northwestward along the deepening seafloor from the equator into the western North Atlantic.

[12] As at 52°W, subtracting 1985 θs from 2003 values along 66°W shows warming within the abyssal thermocline and the nearly homogenous bottom layer below (Figure 3a). This warming is strongest south of 22°N around 5200 dbar. The warming within the AABW-influenced waters for 2003–1985 is ~0.02°C from 18–27°N (Figure 3b). Within the abyssal thermocline, warming occurred prior to 1997 while bottom layer warming is post 1997 and significant at 95%.

[13] The 24°N section data reveal a classic signature of northward flowing AABW [*Wright*, 1970]. A strong zonal gradient in θ is present for P > 4000 dbar (θ < 2.0°C), with potential isotherms overall rising to the east over the western flank of the mid-ocean ridge (Figure 4a). With an interior mid-depth zero-velocity surface, application of the geostrophic relation to the mean density structure yields increasingly northward flow toward the bottom below these sloping isotherms. In contrast, near the western boundary, isotherms rise westward, a signature of the southward-flowing deep-western boundary-current of NADW below ~1000 dbar [*Bryden et al.*, 2005]. Also, isotherms plunge downward toward the ridge within ~1000 dbar of the bottom, likely the result of mixing over the complex ridge topography [*Mauritzen et al.*, 2002].

[14] Subtracting 1981 θs from 2004 values at 24°N reveals a basin-scale pattern (Figure 4a). Warming is evident east of ~57°W, and cooling west of that longitude. This pattern is caused by isotherms deepening east of 57°W, and shoaling to the west. It is consistent with a reduced net northward volume transport of AABW in 2004 versus 1981. To quantify this reduction, data from four sections occupied along 24°N are used to calculate geostrophic meridional velocities employing a 3200-dbar zero-velocity surface [*Bryden et al.*, 2005]. Volume transport integrations are limited to the western basin interior (70–46°W) to exclude the deep-western boundary-current, and to water with θ < 1.8°C to isolate the AABW [*Bryden et al.*, 2005]. North-

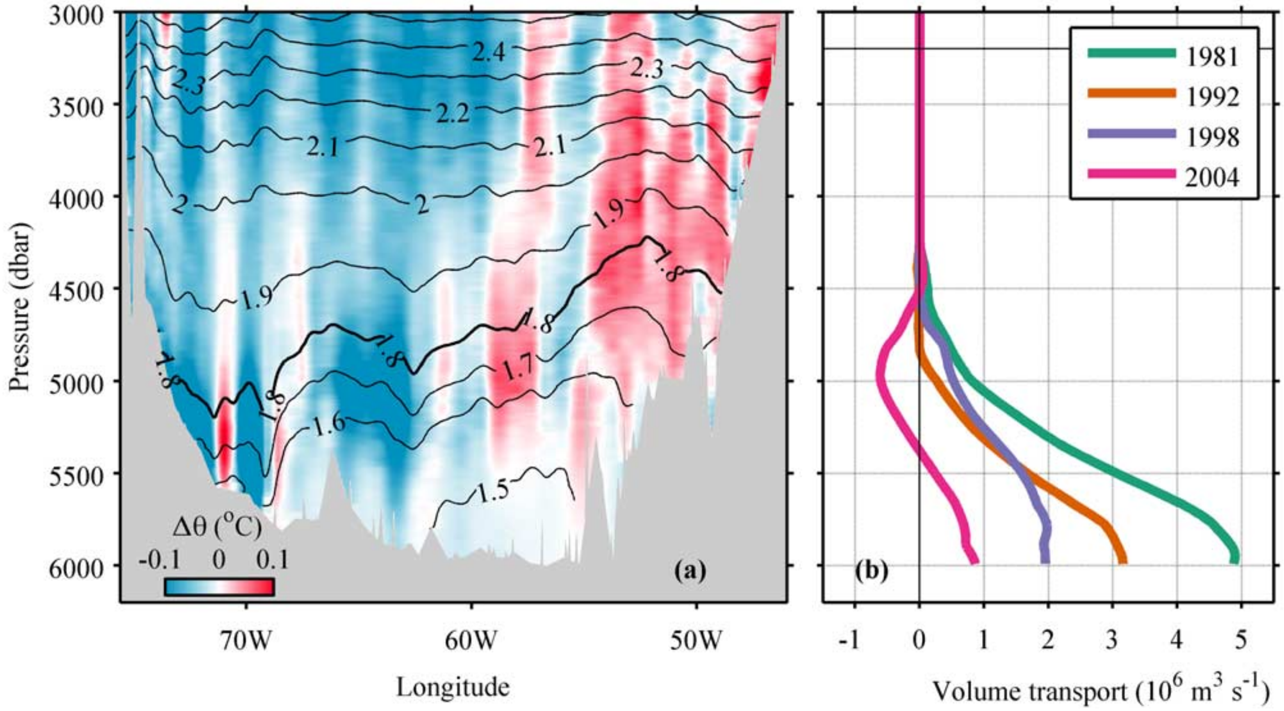

Figure 4. Deep θ and volume transport variability across the western basin of the North Atlantic at 24°N (Figure 1). (a) Deep θ differences (color shading) for 2004–1981 contoured versus pressure and longitude with mean potential isotherms (black contours) from 1981, 1992, 1998, and 2004 data overlaid and bottom bathymetry (grey shading) indicated. (b) Cumulative downward vertical integral of meridional geostrophic volume transport referenced to a 3200-dbar zero-velocity surface for $\theta < 1.8$°C and 70–46°W for 1981, 1992, 1998, and 2004 (see legend).

ward volume transport of AABW estimated from the sections across 24°N decreases monotonically from 1981–2004 (Figure 4b).

4. Discussion

[15] The finding of a northward transport reduction of AABW at 24°N merits some caveats. Data from an instrument array monitoring the Atlantic MOC reveals transport variations exceeding 30% during a single year [*Cunningham et al.*, 2007], suggesting a decadal 30% MOC reduction inferred from analysis of trans-Atlantic sections occupied along 24°N [*Bryden et al.*, 2005] could be aliased short timescale variability. In addition, although previously employed [*Bryden et al.*, 2005], the 3200-dbar zero-velocity surface is overly simplistic. The circulation in the deep-western boundary-current (while excluded here) is also complex and strongly time dependent [*Cunningham et al.*, 2007]. Nonetheless, the large-scale changes observed at 24°N are consistent with a monotonic reduction in the large-scale northward flow of AABW there from 1981–2004.

[16] A 4×10^6 m^3 s^{-1} reduction in northward transport of AABW into the North Atlantic applied steadily over 23 years yields a volume reduction of 1.5×10^{15} m^3. The climatological volume [*Gouretski and Koltermann*, 2004] of water in the western basins of the North Atlantic from 0–48°N with $\theta < 1.8$°C is 4.0×10^{15} m^3, with average thickness of 690 m over an area of 5.8×10^{12} m^2. Thus the inferred volume reduction of AABW is 37% of its climatological value, broadly consistent with the warming reported here originating from some combination of isotherm deepening and lateral retraction within the AABW.

[17] This analysis reveals various signatures of AABW retreat in the western basins of the North Atlantic. At the southern ends of both the 52°W and 66°W sections, the abyssal thermocline between NADW and AABW deepens from the early 1980s to 2003. This deepening results in a reduction of AABW volume there. In addition, the 500-m thick layer of AABW found below that thermocline at 66°W exhibits a warming of ~0.02°C from 1997 to 2003, significant at 95%, another signature of reduced AABW influence. At both longitudes, the 1997 values are mostly intermediate between those of the earlier and later sections. Finally, estimated northward volume transports of AABW across 24°N decrease with time, consistent with a large-scale reduction in the zonal temperature gradient across the deep portions of the western basin.

[18] Our findings complement reports of AABW warming or volume reductions over the past few decades from its Southern Ocean origins to the equatorial Atlantic (Section 1). In contrast, cooling of the densest NADW constituent in the northern North Atlantic in recent decades, along with rapid downstream attenuation of NADW property variations [*Yashayaev and Dickson*, 2008], suggest that NADW may not play a strong role in the AABW warming. However, the abyssal thermocline deepening at the southern ends of both the 52°W and 66°W sections might be linked to other Atlantic MOC variations.

[19] AABW warming reduces its density [*Jacobs et al.*, 2002], and perhaps its formation rate. The AABW warming may be as geographically widespread as AABW itself, which covers much of the global ocean floor and often extends over 1000s of meters in the vertical [*Johnson*, 2008]. Thus, AABW warming over the past decade should contribute to global heat and sea level budgets, helping to close recently reported multi-decadal [*Domingues et al.*, 2008], and perhaps interannual [*Willis et al.*, 2008] imbalances among sea level, mass, and upper ocean heat budgets.

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