Analyzing Changes in the Atlantic Meridional Overturning Circulation Using Oceanic Oxygen and Chlorofluorocarbon Observations

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Abstract

Observations and climate model simulations suggest that the Atlantic Meridional Overturning Circulation (AMOC) may weaken in response to anthropogenic climate forcing. Detecting an anthropogenic AMOC weakening has proven difficult because direct AMOC observations are sparse and have a relatively low signal-to-noise ratio (SNR). Previous studies hypothesize that analyzing hydrographic tracers that are mechanistically linked to AMOC changes may help to increase the SNR. Here we analyze published apparent oxygen utilization (AOU) and chlorofluorocarbons (CFC) observations at 25°N in the Atlantic Ocean spanning the last three decades. We estimate the nonparametric probability density functions of AOU concentrations using Bayesian kriging. The estimated AOU concentrations have decreased, which is inconsistent with predictions from several model simulations that show an AMOC weakening. CFC observations, however, suggest an AMOC weakening. We conclude that the natural variability of these observations may at this time be too large to detect a statistically significant anthropogenic AMOC weakening.

Keywords: Atlantic Meridional Overturning Circulation, anthropogenic climate change, apparent oxygen utilization, chlorofluorocarbons, NADW, AABW, Bayesian kriging

Introduction

The Atlantic Meridional Overturning Circulation (AMOC) is a key component of the global ocean circulation system that transports surface waters from the tropics to higher latitudes where it sinks and returns southward (Bryden and Imawaki 2001). Paleo-observations show that the AMOC has collapsed in the past, which has resulted in global-scale abrupt climate change (Clark *et al.* 2002). Model projections suggest that the AMOC may weaken or collapse in response to anthropogenic climate forcing in the future (Gregory *et al.* 2005; Meehl *et al.* 2007). An AMOC weakening or collapse may impose sizeable impacts on natural systems and human welfare (Keller *et al.* 2000; Link and Tol 2002; Vellinga and Wood 2002; Schneider *et al.* 2007).

Several studies have analyzed AMOC variability in observations collected over the last five decades. Bryden *et al.* (2005) analyzes density observations from five zonal transects at 25°N and concludes that the AMOC has weakened by 30% from 1957 to 2004. Cunningham and Alderson (2007) reports changes in temperature and salinity at 25°N that are consistent with an AMOC weakening. These studies break important new ground towards determining the effect of anthropogenic activities on the AMOC, but analyze variables with relatively low signal-to-noise ratios (SNR) (Broecker *et al.* 1998; Baehr *et al*. 2008; Brennan *et al.* 2008). Several additional tracers that have been hypothesized to be mechanistically linked to AMOC changes may have a higher SNR. Here we analyze whether these tracers show trends that are consistent with a recent AMOC weakening.

Examining hydrographic tracers that are mechanistically linked to AMOC changes may provide an avenue towards increasing the SNR of potential AMOC trends (Broecker *et al.* 1998; Sutherland *et al.* 2001; Brennan *et al.* 2008). In particular, we focus on two mechanisms that link tracers to AMOC changes. The first mechanism involves the composition of the deepwater at 25°N, a mix of two distinct water masses: (i) North Atlantic Deep Water (NADW) and (ii) the underlying Antarctic Bottom Water (AABW) (Bryden *et al.* 1996). The mixture of these two water types at 25°N depends on the deepwater formation rates in the Northern and Southern Hemispheres. Sutherland *et al.* (2001) hypothesize that a decrease in the NADW formation rate would result in an increased formation rate of AABW at 25°N (Broecker 1998) and a shoaling of the NADW/AABW boundary. Because NADW and AABW have different hydrographic properties (Reid 2005), such a shoaling of the boundary should create a unique spatiotemporal pattern in the tracer fields.

Brennan *et al.* (2008) test this hypothesis within a climate model by analyzing one such hydrographic tracer, apparent oxygen utilization (AOU - the difference between observed oxygen concentrations at saturation and observed oxygen concentration of a particular water parcel). Two potential advantages of analyzing AOU are that (i) AOU is a sensitive indicator of changes in ocean circulation (Bopp *et al.* 2002; Matear and Hirst, 2003) and (ii) AOU can be observed with small observation errors relative to expected trends (Keller *et al.* 2002; Deutsch *et al.* 2006; Levine *et al.* 2008). Brennan *et al.* (2008) shows that some model simulations (e.g. the CSIRO model simulations of Matear and Hirst (2003)) predict a shoaling of the NADW/AABW interface, which is consistent with the hypothesis of Sutherland *et al.* (2001)**.** Furthermore, the study shows that analyzing AOU observations at 25°N enables an earlier detection of AMOC changes in the considered model simulation than analyzing AMOC reconstructions based on temperature and salinity observations.

A second mechanism that links hydrographic tracers to AMOC changes involves the depth of the NADW cores in the North Atlantic (Drijfhout and Hazeleger 2007). At 25°N, NADW consists primarily of two different water masses, upper and lower NADW (Dickson and Brown 1994). The depth of these two cores is closely related to the density at their formation sites (Reid 1995). It is hypothesized that if the AMOC were to weaken due to a freshening or warming of surface waters in the North Atlantic, then the densities at the formation sites, and hence, the depth of the upper and lower NADW cores, will decrease. Results of model simulations, indeed, show such a positive correlation between a shoaling of the NADW cores and an AMOC weakening (Drijfhout and Hazeleger 2007). The shoaling of these cores might be evident in chlorofluorocarbons (CFCs) observations because the maxima subsurface CFC concentrations at 25°N occur in the NADW cores and because CFCs can be observed with high precision (Smethie and Fine 2001).

Determining whether observed AOU and CFC trends are consistent with these predictions is challenging for several reasons. First, AOU and CFC observations are sparse in space and time (Cunningham *et al.* 2005). Quantifying potential AOU and CFC trends, hence, requires interpolation. Second, observed AOU trends are subject to considerable uncertainty from internal variability (Keller *et al*. 2002; Cunningham *et al.* 2005). Hence, quantitatively determining whether the model predictions are consistent with the AOU observations requires deriving the full probability density function of AOU concentrations.

Here we analyze AOU and CFC observations spanning the last three decades to evaluate the evidence for a shoaling of (i) the NADW/AABW boundary and (ii) the upper and lower NADW cores. Specifically, we use Bayesian kriging to estimate the nonparametric probability density function of the change in AOU at the NADW and AABW interface. We also perform a simple analysis of changes in concentrations of CFCs in the deepwaters at 25°N. In particular, we calculate the depths of the upper and lower cores of NADW (Bryden *et al.* 1996) as observed in the CFC11 spatial field. We show that AOU has decreased from 1981 to 2004 at the NADW and AABW interface at 25°N, which is inconsistent with previous model and theoretical predictions (Broecker *et al.* 1998; Sutherland *et al.* 2001; Matear and Hirst 2003). The observed CFC11 concentrations suggest a small shoaling of both NADW cores, which is consistent with model predictions (Drijfhout and Hazeleger 2007). This shoaling, however, is subject to considerable noise. These results suggest that the internal and natural variability of the analyzed AOU and CFC observations may be at this time too large to detect statistically significant AMOC changes via a shoaling of the NADW and AABW boundary or a shoaling of upper and lower NADW cores.

Data

We analyze observations from four transects collected around 25°N over the last three decades. Transects were collected in 1981 (Roemmich and Wunsch 1985), 1992 (Parrilla *et al.* 1994), 1998 (Baringer and Molinari 1999), and 2004 (Cunningham *et al.* 2005) along 24.5°N from the African coast (~15°W) to the Bahama Islands (~75°W) (different latitudes were used at the margins of the 1981, 1998, and 2004 transects due to diplomatic issues). We do not analyze oxygen observations from 1957 to avoid complications due to a change in the oxygen measurement method (Bryden *et al.* 1996). We remove all flagged data points, outliers (*cf.* Gouretski and Jancke 1999), and duplicate observations at the same location. We calculate AOU concentrations from observations of oxygen, potential temperature and salinity (*cf.* Weiss 1970). CFC11 concentration observations are only analyzed for the 1992, 1998 and 2004 transects due to limited data availability in the 1957 and 1981 transects.

Methods

We estimate the change in AOU concentration at the interface between NADW and AABW. Specifically, we analyze data from 45-70°W and 3500-6000m, a region surrounding the NADW and AABW boundary (Bryden *et al.* 1996). We first interpolate the observations onto regularly spaced grids using Bayesian kriging (Handcock and Stein 1993). We use Bayesian kriging because it appropriately incorporates the uncertainty in the estimation of the spatial covariance function when estimating uncertainties at each prediction location. Specifically, we use a Markov Chain Monte Carlo (MCMC) algorithm to estimate the exponential covariance and the linear mean function parameters. We use the spBayes package (Finley *et al.* 2007) in the statistical software R

(Ihaka and Gentleman 1996). We account for anisotropy using the 1:100 depth to longitude ratio calculated by Roemmich (1983). We then use the covariance parameters to predict AOU concentrations at a 2.5° longitude and 250 meter depth resolution.

We analyze CFC observations by estimating the changes in the locations of the two NADW cores. We first interpolate the observations onto a 1° longitude by 50 meter grid using cubic splines. We assume that the two maxima of CFC concentrations near the western boundary are the depths of upper and lower NADW branches (Bryden *et al.* 1996). We estimate the depth of the maximum concentration within each core at 71°W for the 1992, 1998 and 2004 transects.

Results and Discussion

AOU increases both with depth and to the east in all four transects (Figure 1). This pattern is consistent with NADW overlying AABW (*cf.* Bryden *et al.* 1996). Difference maps of AOU concentrations since 1981 show mostly decreases with the exception of small increases in 1992 (Figure 1). From 1981 to 1992, AOU decreases slightly in the upper half of the region, but increases near the base of the Mid-Atlantic Ridge. AOU decreases moderately throughout the region from 1981 to 1998, however, with the largest changes in the shallowest regions. From 1981 to 2004, the decrease in AOU is even larger.

Mean AOU concentrations have mostly decreased over the last three decades (Figure 2). Although mean AOU concentration increases slightly from 77 µmol/kg in 1981 to 80 μmol/kg in 1992, it decreases sharply to 73 μmol/kg in 1998 and 70 μmol/kg in 2004. The overall decrease in AOU concentration from 1981 to 2004 is 7 μmol/kg. The widths of the 90% confidence intervals of mean AOU concentration for the 1981, 1992, and 1998 transects are 2.1, 1.8, and 1.9 μmol/kg, respectively. The 90% confidence interval for the 2004 transect, however, is slightly larger at 2.8 μmol/kg. We hypothesize that the larger uncertainty in the 2004 confidence interval results from complications associated with the oxygen measurements during the 2004 cruise (Cunningham *et al.* 2005).

The AOU trends are inconsistent with a shoaling of the boundary between NADW and AABW as hypothesized by Broecker *et al.* (1998) and Sutherland *et al.* (2001) and seen in a model simulation of the CSIRO model (Matear and Hirst 2003). This discrepancy may indicate (i) large uncertainties due to internal variability and observation error (*cf.* Brennen *et al.* 2008), (ii) a confounding of the mechanistic link between AOU and the AMOC by other factors, (iii) the hypothesis of a shoaling of the boundary between NADW and AABW in response to an AMOC weakening has to be rejected, or (iv) the AMOC did not weaken over the considered time interval. More detailed analyses (e.g., via an assimilation of tracers into an Earth System Model [Schmittner *et al*. 2009]) may be able to provide more detailed insights into this question, but are beyond the scope of this analysis.

The AOU trends, however, are consistent with trends of observed temperature and volume of AABW in the North Atlantic over the last three decades (Johnson *et al.* 2008). They report a warming of AABW at 25°N resulting from a deepening of the strong thermal boundary between NADW and AABW, reducing the relative volume of AABW. The observed decrease in AOU in this region is consistent with a reduction in the relative contribution of AABW at the analyzed location.

Our analysis of CFC11 concentrations (Figure 3a, only 1998 is shown) shows an overall increase in CFC11 concentrations throughout the analyzed region, as well as changes in the depths of the NADW cores. In particular, CFC11 increases in the surface waters, as well as in the upper and lower NADW cores. The depths of both NADW cores increase slightly from 1992 to 1998, but decrease more drastically from 1998 to 2004 (Figure 3b). The overall trend from 1992 to 2004 suggests a shoaling of both cores. We note that estimating trends from three data points is difficult, but these results are consistent with the hypothesis that the formation of NADW has decreased from 1992 to 2004 due to an AMOC weakening (Drijfhout and Hazeleger 2007).

Caveats

Our results hinge on several key assumptions. First, we infer changes in the AMOC from the analysis of four zonal transects at 25°N in the North Atlantic from 1981 to 2004. The AMOC has been shown to be highly variable on decadal to centennial timescales, limiting our ability to determine secular trends with confidence (Wunsch 2008). Second, we analyze the spatial and temporal variability of AOU and CFC11 concentrations as a proxy for changes in deepwater formation rates. Observed changes may, instead, reflect variability unrelated to changes in deepwater formation rates.

Conclusions

Given our aforementioned assumptions and caveats, we draw two main conclusions. First, we find a decrease in AOU concentrations from 1981 to 2004 in the deepwaters of the western Atlantic at 25°N. This trend is inconsistent with the shoaling of the boundary between NADW and AABW as hypothesized by, for example, Sutherland *et al.* (2001). Second, we show that the cores of CFC11 concentrations in this region shoal from 1992 to 2004, consistent with the hypothesis of a reduction of NADW formation and an AMOC weakening. Additionally, we find that the 90% confidence intervals of the mean AOU concentrations at the NADW and AABW interface are small relative to observed changes. Assimilating these AOU observations into model simulations, hence, may provide a potentially useful avenue to improve AMOC projections.

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References

Baehr, J., K. Keller and J. Marotzke (2008), Detecting potential changes in the meridional overturning circulation at 26°N in the Atlantic, *Climatic Change*, *91*, 11-27.

Baringer M.O. and Molinari (1999), Atlantic Ocean baroclinic heat flux at 24° to 26°, *Geophysical Research Letters*, *26*, 353-356.

Bopp, L., C. Le Quéré, M. Heimann, A. C. Manning, and P. Monfray (2002), Climate-induced oceanic oxygen fluxes: Implications for the contemporary carbon budget, *Global Biogeochemical Cycles*, *16*, doi 10.1029/2001GB001445.

Brennan, C. E., R. Matear, and K. Keller (2008), Measuring oxygen concentrations to improve the detection capabilities of an ocean circulation observation array, *Journal of Geophysical Research-Oceans*, *113*, C02019, doi:10.1029/2007JC004113.

Broecker, W. S. (1998), Paleocean circulation during the last deglaciation: A bipolar seesaw?, *Paleoceanography*, *13*, 119-121.

Broecker, W. S., S. L. Peacock, S. Walker, R. Weiss, E. Fahrbach, M. Schroeder, U. Mikolajewicz, C. Heinze, R. Key, T.-H. Peng, and S. Rubin (1998), How much deep water is formed in the Southern Ocean?, *Journal of Geophysical Research-Oceans*, *103*, 15833-15843.

Bryden, H. L., M. J. Griffiths, A. M. Lavin, R. C. Millard, G. Parrilla, and W. M. Smethie (1996), Decadal changes in water mass characteristics at 24 degrees N in the subtropical North Atlantic ocean, *Journal of Climate*, *9*, 3162-3186.

Bryden, H. L., H. R. Longworth, and S. A. Cunningham (2005), Slowing of the Atlantic meridional overturning circulation at 25 degrees N, *Nature*, *438*, 655-657.

Clark, P. U., N. G. Pisias, T. F. Stocker, and A. J. Weaver, (2002) The role of the thermohaline circulation in abrupt climate change. *Nature*, *415***,** 863-869.

Cunningham S.A. and S. Alderson, (2007) Transatlantic temperature and salinity changes at 24.5°N from 1957 to 2004. *Geophysical Research Letters, 34,* L14606, doi:10.1029/2007GL029821.

Cunningham S.A. *et al.*, (2005), RRS *Discovery* Cruise D279, 04 Apr – 10 May 2004. A Transatlantic hydrographic section at 24.5°N. *Southampton Oceanography Centre Cruise Report*, 198pp.

Deutsch, C., S. Emerson, and L. Thompson (2006), Physical-biological interactions in North Pacific oxygen variability. *Journal of Geophysical Research*, *111*, C09S90, doi:10.1029/2005JC003179.

Dickson, R.R. and J. Brown (1994), The production of North Atlantic Deep Water: Sources, rates, and pathways. *Journal of Geophysical Research, 99*, 12319-12341.

Drijfhout, S.S. and W. Hazeleger (2007), Detecting Atlantic MOC changes in an ensemble of climate change simulations, *Journal of Climate*, 20, 1571-1582.

Finley, A. O., S. Banerjee, and B. P. Carlin (2007), spBayes: A program for multivariate point-referenced spatial modeling, *Journal of Statistical Software*, *19*.

Gouretski, V. V., and J. K. (1999), A Description and Quality Assessment of the Historical Hydrographic Data for the South Pacific Ocean, *Journal of Atmospheric and Oceanic Technology*, *16*, 1791-1815.

Gregory, J. M., K. W. Dixon, R. J. Stouffer, A. J. Weaver, E. Driesschaert, M. Eby, T. Fichefet, H. Hasumi, A. Hu, J. H. Jungclaus, I. V. Kamenkovich, A. Levermann, M. Montoya, S. Murakami, S. Nawrath, A. Oka, A. P. Sokolov, and R. B. Thorpe (2005), A model intercomparison of changes in the Atlantic thermohaline circulation in response to increasing atmospheric CO_2 concentration, *Geophysical Research Letters*, *32,* L12703, doi:10.1029/2005GL023209.

Handcock, M.S. and M.L. Stein (1993), A Bayesian analysis of kriging, *Technometrics*, *35*, 403-410.

Ihaka, R. and R. Gentleman (1996), R: A language for data analysis and graphics, *Journal of Computational and Graphical Statistics*, *5*, 299-314.

Johnson, G. C., S. G. Purkey, and J. M. Toole (2008), Reduced Antarctic meridional overturning circulation reaches the North Atlantic Ocean, *Geophysical Research Letters*, *35*, L22601, doi:10.1029/2008GL035619.

Keller, K., R. D. Slater, M. Bender and R. M. Key (2002), Possible biological or physical explanations for decadal scale trends in North Pacific nutrient concentrations and oxygen utilization, *Deep-Sea-Research* II, 49, 345-362.

Keller, K., K. Tan, F. M. M. Morel, and D. F. Bradford (2000), Preserving the ocean circulation: Implications for climate policy, *Climatic Change*, 47, 17-43.

Levine, N. M., S. C. Doney, R. Wanninkhof, and K. Lindsay (2008), Impact of ocean carbon system variability on the detection of temporal increases in anthropogenic CO_2, *Journal of Geophysical Research*, *113*, C03019, doi:10.1029/2007JC004153.

Link, P. M. and R. S. J. Tol (2004) Possible Economic Impacts of a Slowdown of the Thermohaline Circulation: An Application of Fund, *Portuguese Economic Journal,* 3**,** 99.

Matear, R. J., and A. C. Hirst (2003), Long-term changes in dissolved oxygen concentrations in the ocean caused by protracted global warming, *Global Biogeochemical Cycles*, *17*, 1125, doi:10.1029/2002GB001997.

Meehl, G.A., T.F stocker, W.D. Collins, P. Friedlingstein, A.T. Gaye, J.M. Gregory, A. Kitch, R. Knutti, J.M. Murphy, A. Noda, S.C.B. Raper, I.G. Watterson, A.J. Weaver and Z.-C. Zhao (2007), Global Climate Projections, *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller, Eds, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Parrilla R., A. Lavin, H. Bryden, M. Garcia and R. Millard, (1994), Rising temperatures in the subtropical North Atlantic Ocean over the past 35 years, *Nature, 369*, 48-51.

Reid, J.L. (2005), On the world-wide circulation of the deep water from the North Atlantic Ocean, *Journal of Marine Research*, *63*, 187-201.

Roemmich, D. (1983), Optimal estimation of hydrographic station data and derived fields, *Journal of Physical Oceanography*, *13*, 1544-1548.

Roemmich, D. and C. Wunsch (1985), Two transatlantic sections: meridional circulation and heat flux in the subtropical North Atlantic Ocean, *Deep-Sea Research*, *32*, 619-664.

Schmittner, A., N.M. Urban, K. Keller and D. Matthews (2009), Using tracer observations to reduce the uncertainty of ocean diapycnal mixing and climate-carbon cycle projections, *Global Biogeochemical Cycles*, (in review, e-print available at http://www.geosc.psu.edu/~kkeller/Schmittner_et_al_gbc_08.pdf).

Schneider, S.H., S. Semenov, A. Patwardhan, I. Burton, C.H.D. Magadza, M. Oppenheimer, A.B. Pittock, A. Rahman, J.B. Smith, A. Suarez, and F. Yamin (2007), Assessing key vulnerabilities and the risk from climate change, *Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, M.L. Parry, O.F. Canziani, J.P. Palutikof, P.J. van der Linden and C.E. Hanson., Eds., Cambridge University, UK, 779-810.

Smethie Jr., W.M. and R.A. Fine (2001), Rates of North Atlantic Deep Water formation calculated from chlorofluorocarbon inventories, *Deep-Sea Research I*, 48, 189-215.

Sutherland, S. C., W. S. Broecker, and T. Takahashi (2001), Stability of the boundary separating Antarctic Bottom Water from North Atlantic Deep Water in the western South Atlantic, *Geophysical Research Letters*, *28*, 4219-4222.

Vellinga, M., and R. A. Wood (2002), Global climatic impacts of a collapse of the Atlantic thermohaline circulation, *Climatic Change*, *54*, 251-267.

Weiss, R. F. (1970) The solubility of nitrogen, oxygen and argon in water and seawater. *Deep Sea Research*, *17*, 712–735.

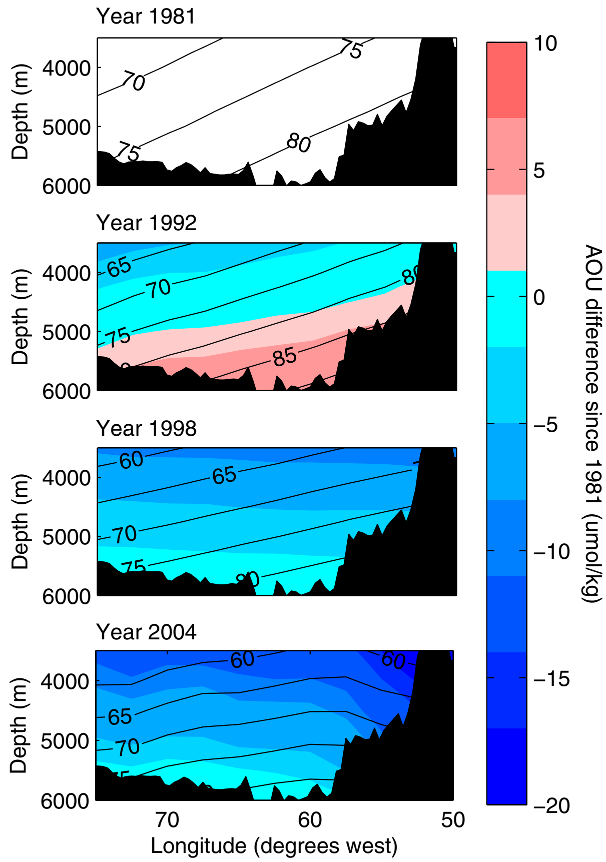
Wunsch, C. (2008), Mass and Volume Transport Variability in an Eddy-Filled Ocean, *Nature Geoscience*, *1*, 165-168.

List of Figure Captions

Figure 1. Contour maps of interpolated AOU concentrations in µmol/kg (contour lines) (Roemmich and Wunsch 1985; Parrilla *et al.* 1994; Baringer and Molinari 1999; Cunningham *et al.* 2005) and temporal trends relative to the 1981 transect (shaded).

Figure 2. Mean AOU concentration in the considered region of the transect. The line within each box shows the median and the edges of each box mark the 25% and 75% confidence limits of the mean AOU concentrations. The error bars represent the 5% and 95% confidence limits derived from the posterior distributions in the Bayesian kriging analysis.

Figure 3. (a) Contour map of CFC11 concentrations for the 1998 transect (pmol/kg) (Baringer and Molinari 1999). The 'X' symbols mark the upper and lower cores of NADW. (b) The depths of the maximum CFC11 concentration in the upper and lower branches of NADW at 71°W for the 1992 (Parrilla *et al.* 1994), 1998 (Baringer and Molinari 1999) and 2004 transects (Cunningham *et al.* 2005). The trend lines shown are derived from linear least-squares regression.



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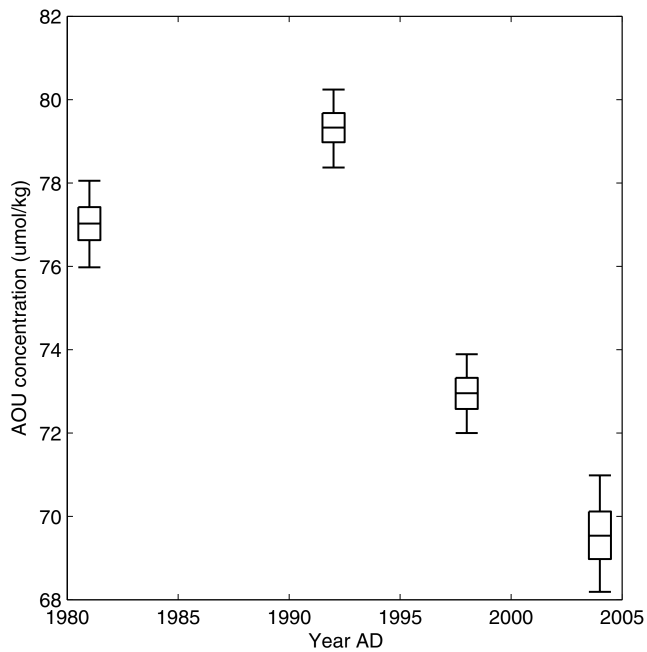

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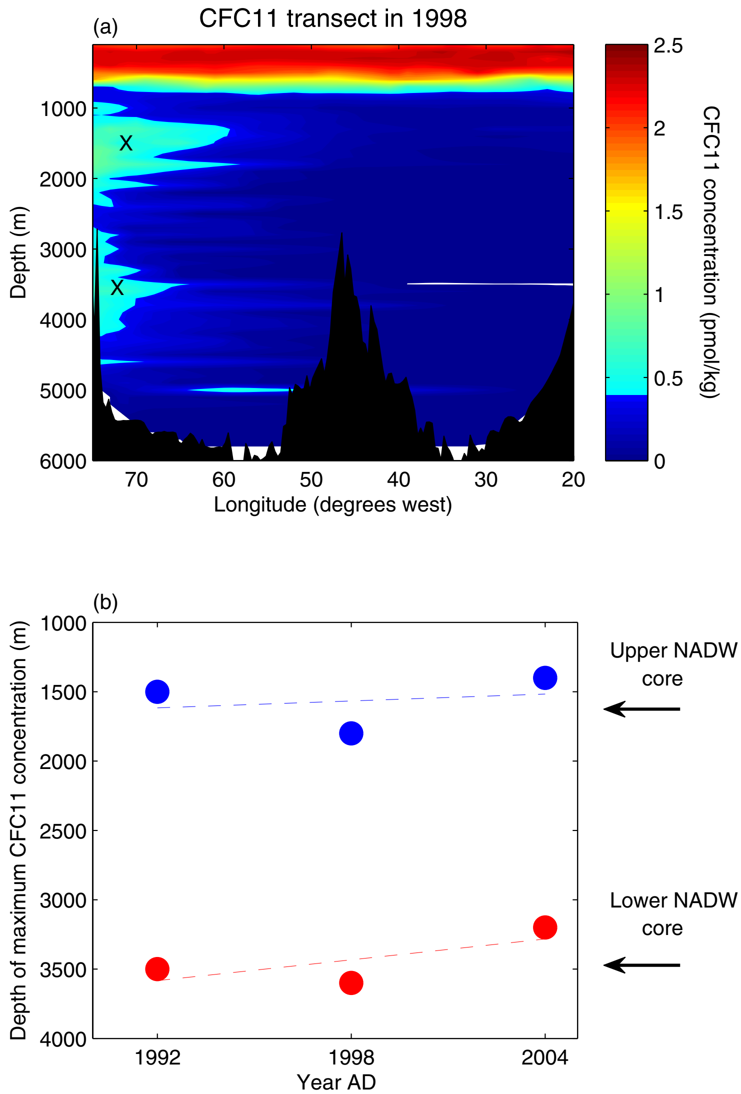

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