Is the Thermohaline Circulation Changing?

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Submitted to J. Climate, final version, 21st December 2005

Abstract

Analyses of ocean observations and model simulations suggest considerable changes in the thermohaline circulation (THC) during the last century. These changes are likely to be the result of natural multidecadal climate variability and driven by low-frequency variations of the North Atlantic Oscillation (NAO) through changes in Labrador Sea convection. Indications of a sustained THC weakening are not seen during the last few decades. Instead a strengthening since the 1980s is observed. The combined assessment of ocean hydrography data and model results indicates that the expected anthropogenic weakening of the THC will remain within the range of natural variability during the next several decades.

1. Introduction

The thermohaline circulation (THC) is a global 3-dimensional belt of ocean currents that transports large amounts of heat and freshwater (Manabe and Stouffer, 1999). In the North Atlantic, it is manifested in a meridional overturning circulation (MOC), which, through its northward transport of warm tropical waters by the Gulf Stream and North Atlantic Current, effectively contributes to the warming of Northern Europe. It has been proposed by Manabe et al. (1991) and others that global warming may lead to a substantial weakening of the MOC. On its own this could have serious impacts on the climate, the ecology and the economy of many countries surrounding the North Atlantic. However, the cooling associated with the THC weakening compensates partly the greenhouse warming in the North Atlantic, thereby competing with the effect of global climate change in this region.

Direct measurements of the MOC are rare, so that it has proven difficult to quantitatively derive its low-frequency variability (Lumpkin et al., submitted 2005). One possible way to infer the relative strength of the MOC is the use of sea surface temperature (SST) observations which are available for the last century (Rayner et al., 2003). As shown by a host of global climate model studies (e.g., Manabe and Stouffer, 1988; Vellinga and Wood, 2002), variations of the MOC on multidecadal and longer timescales are accompanied by a characteristic interhemispheric SST anomaly pattern, with anomalies of opposite signs in the North and South Atlantic. This dipolar SST anomaly pattern in the Atlantic has been recognized as one of the main multidecadal variability modes during the last century (Folland et al., 1986; Mestaz-Nunez and Enfield, 1999; Knight et al., 2005). The dipolar SST anomaly pattern is markedly different to that prevailing on interannual to decadal timescales which is tripolar in the North Atlantic (Visbeck et al., 1998). As an example of the multidecadal variability, the linear trend in the global SST observed during the period 1980-2004 is displayed in Fig. 1. The global mean trend was removed to highlight the dynamical changes. Two features stand out: A Pacific SST anomaly pattern resembling the Pacific Decadal

Oscillation (PDO) (Mantua et al., 1997) and the dipolar interhemispheric SST anomaly pattern in the Atlantic.

2. Causes of MOC changes

In many climate change simulations, the strength of the MOC has been found to be tied to the surface to mid-depth density gradient between the subpolar North Atlantic and the South Atlantic (Hughes and Weaver, 1994; Thorpe et al., 2001; Rahmstorf, 1996). The density of the deep waters in the subarctic Atlantic that comprise the lower, southward limb of the MOC, is set by two main contributions: The near-bottom outflow of waters from the Nordic Seas spilling across the Greenland-Iceland-Scotland ridge, and the overlying waters formed during episodes of strong wintertime cooling in the Labrador Sea (LSW). The variability of formation of the Labrador Sea Water is mainly driven by the North Atlantic Oscillation (NAO), a leading mode of natural atmospheric variability (Hurrell, 1995). Changes in the NAO index (Fig. 2), a measure of the strength of the westerlies and heat fluxes over the North Atlantic, influence deep convection in the Labrador Sea and are reflected in the properties of the LSW (Fig. 2) (Curry et al., 1998). Coupled model integrations by Delworth and Dixon, 2000 show that an anthropogenically induced slowdown of the MOC may be delayed by several decades in response to a sustained upward trend of the NAO during winter, such as has been observed over the last 30 years.

Ocean model studies indicate that NAO-related variations in the heat fluxes over the Labrador Sea induce a lagged (2-3 years) response of the MOC (Eden and Willebrand, 2001; Häkkinen, 1999) that is rapidly communicated between subpolar and tropical latitudes through boundary wave processes (Getzlaff et al., 2005). On the longer multidecadal time scales, the corresponding changes in the interhemispheric transport of heat induce the above described dipole SST anomaly pattern (Visbeck et al., 1998), involving SST anomalies particularly along the North Atlantic Current that develop against the local heat fluxes

(Bjerknes, 1964; Kushnir; 1994; Eden and Jung, 2001). On these multidecadal timescales, the dipole SST anomaly pattern can thus be used as a fingerprint to detect changes in the MOC (Latif et al., 2004).

An Atlantic dipole SST anomaly index was computed from 1900 onwards using observations (Rayner et al., 2003), as the difference of the annual mean SSTs averaged over the regions 40-60^o N, 60-10^{o}W and 40-60^{o}S, 50-0^{o}W, two regions of strong multidecadal changes in the North and South Atlantic during the most recent decades (Fig. 1). The dipole index exhibits pronounced multidecadal changes (Fig. 2), with a marked decline from the 1920s to the 1970s, and a recovery thereafter. A strong downward trend that may indicate an anthropogenic weakening of the MOC is not obvious. Using the coupled model results of Latif et al., 2004 a change of 1°C in the dipole index would correspond to about a 3 Sv (1Sv=$10^6 m^3/s$) change of the MOC. The magnitude of typical decadal-scale anomalies in the observed dipole SST anomaly index amount to about 0.5-1.0°C, which would yield anomalies in the strength of the MOC of the order of 1.5 to 3 Sv, if the model relationship carries over into the real world.

Clearly, the strong low-frequency variations of the dipole SST anomaly index lag the corresponding variations of the NAO by roughly a decade. A cross correlation analysis between the two indices for the time period 1900-2004 supports the visual impression (Fig. 3). The structure of the cross correlation function is consistent with the notion that the NAO drives the variations in the SST dipole, with correlations of about 0.2 to 0.3 at lags between -5 and -20 years (negative lags indicate that the NAO leads). Given the large number of degrees of freedom in the NAO time series, correlations of 0.2 are significant at the 95% level. It is thus plausible to interpret the multidecadal SST variations as being induced by multidecadal changes of the MOC in response to the low-frequency atmospheric forcing associated with the

NAO and the associated changes in Labrador Sea convection, a result which has been shown so far only by model simulations (Eden and Jung, 2001).

A decadal-scale trend of the MOC consistent with the inference drawn from the observed SSTs has been a robust feature of several hindcasting studies with a variety of ocean models. Simulations considering the response to realistic atmospheric forcing derived either from ship-based data (da Silva et al., 1994) or atmospheric reanalyses products (Kalnay et al., 1996), typically show an increase of the MOC by about 20% of its mean, from lowest values in the late 1960s or early 1970s to a maximum in the mid-1990s (Häkkinen, 1999; Eden and Jung, 2001; Bentsen et al., 2004; Beismann and Barnier, 2004). This figure is consistent with that derived from the dipole SST anomaly index, as shown above. The simulated changes clearly follow the multidecadal variations in the NAO-index and are attributed primarily to the effect of the heat flux changes over the subpolar North Atlantic affecting Labrador Sea convection. Changes in the simulated hydrographic properties such as the thickness of LSW or indices for the strength of horizontal circulation in the North Atlantic are generally consistent with observations (Bentsen et al., 2004; Haak et al., 2003).

3. The role of the overflows

If the observed multidecadal changes in the interhemispheric SST anomaly pattern can primarily be rationalized in terms of the MOC response to NAO-related variations in the deep water formation in the Labrador Sea, the question arises what the role of changes in the near-bottom outflow of cold waters from the Nordic Seas is. In the present climate, the overflows across the Greenland-Iceland-Scotland ridge system provide the densest source waters to the deep southward branch of the MOC. Model studies suggest that changes in the density of the outflow could potentially induce much larger effects on the MOC than the atmospheric forcing over the subpolar North Atlantic (Döscher and Redler, 1997), and would be a prime agent involved in inducing a significant slackening of the MOC in response to global warming (Wood et al., 1999).

Long-term observational records from the overflows indicate considerable freshening and cooling of the deep waters over the past decades (Dickson et al., 2002; Curry and Mauritzen, 2005). An estimate of the potential effect of changes in the density of the overflows is obtained from simulations with two different ocean models: one global model and one regional Atlantic model (see Appendix for details). A number of sensitivity experiments was performed with the models, which allow an assessment of the response of the MOC strength to changes in the density of the overflow through Denmark Strait on decadal timescales (see Appendix for details). A summary plot with the results of all experiments is given in Fig. 4. The major result is that both models exhibit a linear relationship between the MOC transport and density changes of the Denmark Strait overflow.

The observed salinity change of the deep water at the sill of Denmark Strait from 1970-2000 amounts to approximately 0.03 psu which would correspond to a density change of 0.024 kg/m^3. Temperature observations (Dickson and Meincke, 2003), however, reveal a simultaneous cooling which has a compensating effect on the density of about 50% so that the net density change is about 0.012 kg/m^3. On the basis of the modeling results shown in Fig. 4, this translates to a maximum effect on the MOC on the order of 1 Sv, which is well within the range of the variability associated with the NAO-driven changes in LSW formation during the last century (±1.5-3.0 Sv), as discussed above. It should be noted here that the density of the outflow waters is modified by entrainment of ambient waters in the downslope flow region. The analysis of the effects of possible variations in the entrainment rate on the MOC is, however, beyond the scope of the present paper.

4. Summary and discussion

We have investigated the multidecadal variability of the Atlantic MOC during the 20th century by making use of a characteristic relationship between MOC and SST found in global climate models. The simulated Atlantic SST response to multidecadal changes in the MOC is the interhemispheric dipole pattern, with opposite changes in the North and South Atlantic. An index of the MOC during the 20th century was derived by computing the observed SST difference between the North and South Atlantic. This index exhibits pronounced multidecadal variability during the 20th century. We found evidence that these multidecadal MOC variations can be understood as the lagged response to the multidecadal variations in the NAO and the associated variations in Labrador Sea convection. Finally, we estimated from the analysis of ocean model simulations that the observed density change over the period 1970-2000 in the region of Denmark Straight translates into an MOC change of about 1Sv, which is well within the range of the natural multidecadal variability that we estimated to amount to about ±1.5-3 Sv.

While the combined evidence from surface observations, the hydrographic data and model simulations suggests that the variations in the MOC over the last decades can primarily be regarded as a response to the NAO variability, a continuing freshening (or warming) trend in the Nordic Seas must be considered as a key additional factor for the future evolution of the MOC in view of anthropogenic climate change. A recent assessment of global warming simulations by Schweckendiek and Willebrand, 2005 suggests that 21st-century projections of the MOC are mainly tied to the evolution of hydrographic conditions in the Nordic Seas. Most global climate model projections for the 21st century suggest a gradual anthropogenic weakening of the MOC of up to 40% (IPCC, 2001; Gregory et al., 2005). As described above, the level of internal multidecadal varability has been estimated to about 1.5-3 Sv. Thus, such a weakening will not exceed the range of multidecadal variability within the next several decades.

Our analysis does not provide indications for a sustained weakening of the MOC during the last few decades, which is consistent with the study of Knight et al. 2005. Recently , however, Bryden et al. 2005 found some evidence of such a weakening of the MOC by analyzing a transatlantic section at 25°N that has been occupied five times since 1957. Thus, the current state of the MOC is discussed controversially. As a cautionary remark it is noted that the possibility of increased melting of the Greenland ice sheet, which could substantially change the freshwater budget in the Nordic Seas, is not included in present models. In this regard a monitoring of the tendencies in the hydrographic properties of the overflow should constitute a key element of a long-term observing system for the MOC in the Atlantic Ocean.

Acknowledgements

This work was supported by the Sonderforschungsbereich 460 of the Deutsche Forschungsgemeinschaft and the DYNAMITE and ENSEMBLES projects of the European Union. We thank R. Curry for providing the LSW thickness data, the DRAKKAR group and NEMO system team for close interaction in the set-up of the global model. The ocean model integrations were performed at the German Climate Computing Centre (DKRZ).

Appendix

The global simulations are based on implementations of the recent code version OPA9 of the ORCA-LIM ocean-ice model system described by Madec et al., 1999. Two versions are used here: ORCA2, with a nominal grid size of 2° and ORCA05, with a grid of 0.5° (actual grid sizes in the subpolar North Atlantic are about 120 km and 30 km, respectively). Vertically the water column is divided into 46 (ORCA05) and 31 (ORCA2) levels, with grid cells ranging from 6-10 m at the surface up to 250 m at depth. The representation of bottom topography allows the bottom cell to be filled partially, resulting in a better representation of topographic slopes compared to traditional z-coordinate models. The effect of subgrid-scale eddies on the mixing of tracers is parameterized by rotating the tensor along isopycnals and applying the parameterization of Gent and McWilliams, 1990 (GM90). Vertical eddy viscosity and diffusivity coefficients are calculated by a 1.5 turbulent closure model, static instabilities are removed by increased values. Temperature and salinity are advected using a Monotonic Upstream centered Scheme for Conservation Laws (MUSCL), velocity by conserving total energy for general flow and potential enstrophy for flow with no mass divergence. The global model cases were computed using the forcing fields of Large and Yeager, 2004). The specific implementation is realized via bulk formulae, allowing some feedback of the ocean on the atmospheric fluxes. Changes in overflow density were realized by modifications in the Arctic freshwater budget.

The regional Atlantic simulations utilize a version of the FLAME model hierarchy (Dengg et al., 1999), spanning the Atlantic basin between 70°N and 18°S, with a resolution of about 18 km at 60°N. Vertical resolution increases from 10 m at the surface to 250 m below 2250 m in 45 vertical levels. The code is based on the Modular Ocean Model (MOM2, Pacanowski, 1995), using the bottom boundary layer parameterization of Beckmann and R. Döscher, 1997. Mixing is parameterized by isopycnal diffusion (no GM90 to avoid a reduction of the transport through the Denmark Straits) with a coefficient of 400 m^2/s. Sea ice is not explicitly contained in this model apart from turning off the surface fluxes when SST gets below freezing. Surface boundary layers are implemented by a Kraus-Turner scheme and vertical instabilities are subjected to implicit vertical diffusion. The deep water formation in the Labrador Sea was shown to be in accordance with observations by Böning et al., 2003. Atmospheric forcing is based on climatological ECMWF fields with a mean annual cycle and salinity restoring to monthly mean sea surface salinities. The model is relaxed to observed hydrographic conditions near the closed northern boundary at 70°N. After a spin-up of 22 years, 15-year response experiments were conducted, in which anomalies of (+1°C, 0°C, -1°C) were imposed in the conditions at the northern boundary. This yields changes in the density of the reservoirs north of the sills, and subsequently, of the overflow waters by the amounts shown in Fig. 4. The direct effect of the sponge zone anomalies on the overturning north of the sills is about ±0.3 Sv. An estimate of the upper bound of the MOC variability due to changes in Labrador Sea convection is determined by a dedicated experiment: A complete suppression of convection (by limiting wintertime heat losses) over a 15-year period led to a decline of the MOC of 3.3 Sv (or 22%), consistent with the range of the NAO-related interdecadal variability cited in the text.

References

Bentsen, M., H. Drange, T. Furevik, and T. Zhou, 2004: Simulated varability of the Atlantic meridional overturning circulation. *Clim. Dyn*., **22**, 701-720.

Beismann, J.-O., B. Barnier, 2004: Variability of the meridional overturning circulation of the North Atlantic: sensitivity to overflows of dense water masses. *Ocean Dyn*., **54**, 92-106.

J. Bjerknes, 1964: Atlantic Air-Sea Interactions. *Advances in Geophysics*, **10**, Academic Press, 1-82.

Böning, C.W. , Rhein, M. , Dengg, J., Dorow, C., 2003: Modelling CFC inventories and formation rates of Labrador Sea Water. *Geophys. Res. Lett*., **30** (2), doi: 10.1029/2002GL014855.

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

Curry, R.G., M.S. McCartney, and T.M. Joyce, 1998: Oceanic transport of subpolar climate signals to mid-depth subtropical waters. *Nature*, **391**, 575-577.

Curry,R., C. Mauritzen, 2005: Dilution of the Northern North Atlantic Ocean in Recent Decades. *Science*, **308**, 1772-1774.

Delworth, T.L., K.W. Dixon, 2000: Implications of the recent trend in the Arctic/North Atlantic Oscillation for the North Atlantic thermohaline circulation. *J. Climate*, **13**, 3721-3727.

Dengg, J., Böning, C.W. , Ernst, U. , Redler, R. , Beckmann, A., 1999: Effects of an Improved Model Representation of Overflow Water on the Subpolar North Atlantic. Int. WOCE Newslett., **37**, 10-15.

Dickson, R., Yashayaev, I., Meincke, J., Turrell, B., Dye, S., and Holfort, J., 2002: Rapid freshening of the deep North Atlantic Ocean over the past four decades. *Nature*, **416**, 832-837.

Dickson, R.R., J. Meincke, 2003: The North Atlantic Oscillation and the Ocean's respos in the 1990s. *ICES Mar. Sci. Symp*., **219**, 15-24.

Döscher, R., R. Redler, 1997: The Relative Importance of Northern Overflow and Subpolar Deep Convection for the North Atlantic Thermohaline Circulation. *J. Phys. Oceanogr.*, **27**, 1894-1902.

Eden, C., T. Jung, 2001: North Atlantic Interdecadal Variability: Oceanic response to the North Atlantic Oscillation (1865-1997). *J. Climate*, **14**, 676-691.

Eden, C., J. Willebrand, 2001: Mechanism of interannual to decadal variability of the North Atlantic circulation. *J. Climate*, **14** (10), 2266-2280.

Folland, C.K., T.N. Palmer, and D.E. Parker, 1986: Sahel rainfall and worldwide sea temperature 1901-85. *Nature*, **320**, 602-607.

Gent, P.R., J.C. McWilliams, 1990:Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr*., **20**, 150-155.

Getzlaff, J., C.W. Böning, C. Eden, and A. Biastoch, 2005: Signal propagation related to the North Atlantic overturning. *Geophys. Res. Lett*., **32**, doi:10.1029/2004GL021002.

Gregory, J.M., K. W. Dixon, R. J. Stouffer, *et al.*, 2005: A model intercomparison of changes in the Atlantic thermohaline circulation in response to increasing atmospheric CO_2 concentration. *Geophys. Res. Lett.*, 32, L12703, doi: 10.1029/2005GL023209.

Haak, H., J. Jungclaus, U. Mikolajewicz, and M. Latif, 2003: On the formation and propagation of great salinity anomalies. *Geophys. Res. Lett.*, **30** (9), 1473-76.

Häkkinen, S., 1999: Variability of the simulated meridional heat transport in the North Atlantic for the period 1951-1993. *J. Geophys. Res*., **104** (C5), 10991-11007.

Hughes, T.M.C, A.J. Weaver, 1994: Multiple equilibria of an asymmetric two-basin ocean model. *J. Phys. Oceanogr*., **24**, 619-637.

Hurrell, J.W., 1995: Decadal Trends in the North Atlantic Oscillation: Regional Temperatures and Precipitation. *Science*, **269**, 676-679.

IPCC, 2001: Climate Change. The Scientific Basis. Technical Summary of the Working Group I Report, Cambridge University Press.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, et al., 1996: the NCEP/NCAR 40-Year Reanalyses Project. *Bull. Am. Met. Soc.*, **77**, 437-471.

Knight, J.R., R.J. Allan, C.K. Folland, M. Vellinga, and M.E. Mann, 2005: A signature of persistent natural thermohaline circulation cycles in observed climate. Geophys. Res. Lett., 32, L20708, doi: 10.1029/2005GRL024233.

Kushnir, Y., 1994: Interdecadal Varations in North Atlantic Sea Surface Temperature and Associated Atmospheric Conditions. *J. Climate*, **7** (1), 141-157.

Latif, M., Roeckner, E. , Botzet, M. , et al., 2004: Reconstructing, Monitoring, and Predicting Decadal-Scale Changes in the North Atlantic Thermohaline Circulation with Sea Surface Temperature . *J. Climate*, **17**, 1605-1614.

Large, W.G., S.G. Yeager, 2004: Diurnal to decadal global forcing for ocean and sea-ice models: the data sets and flux climatologies. NCAR Technical Note NCAR/TN-460+STR, NCAR, USA.

Lumpkin, R., K. Speer, and K.P.. Koltermann, 2005: Transport across 48° N in the North Atlantic Ocean. *J. Marine Res.*, submitted.

Madec, G., P. Delecluse, M. Imbard, and C. Levy, 1998: OPA 8.1 Ocean General Circulaton Model. Reference Manual. Technical report, **1**, Laboratoire d'Oceanographie Dynamique et Climatologie, Institut Pierre Simon Laplace des Sciences de l'Environnement Global.

Manabe, S., R.J. Stouffer, 1988: Two Stable Eqilibria of a Coupled Ocean-Atmosphere Model. *J. Climate*, **1**, 841-866.

Manabe, S., R.J. Stouffer, 1999: The role of thermohaline circulation in climate. *Tellus*, **51** (A), 91-109.

Manabe, S., R.J. Stouffer, M.J. Spelman, and K.Bryan, 1991: Transient response of a coupled ocean-atmosphere model to gradual changes of atmospheric CO_2. Part I: Annual mean response. *J. Climate*, **4**, 785-818.

Mantua, N.J., S.R.Hare, Y. Zhang J.M. Wallace, and R.C. Francis, 1997: A Pacific Interdecadal Climate Oscillation with Impacts on Salmon Production. *Bull. Am. Met. Soc.*, **78**, 1069-1079.

Mestaz-Nuñez, A., D.B. Enfield, 1999: Rotated Global Modes of Non-ENSO Sea Surface Temperature Variability. *J. Climate*, **12** (9), 2734-2746.

Pacanowski, R., 1995: User's Guide and Reference Manual. Technical Report 3, GFDL Ocean Group. Available from GFDL, Princeton, New Jersey.

Rahmstorf, S., 1996: On the freshwater forcing and transport of the Atlantic thermohaline circulation. *Clim. Dyn*., **12**, 799-811.

Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res*., **108** (D14), 4407, doi:10.1029/2002JD002670.

Schweckendiek, U., J. Willebrand, 2005: Mechanisms affecting the Overturning Response in Global Warming Simulations. *J. Climate*, **18**, 4925-4936.

da Silva, A., A.C.Young, and S. Levitus, 1994: *Atlas of surface marine data 1994. Volume 1: Algorithms and procedures.* Tech. Rep., **6**, U.S. Department of Commerce, NOAA, NESDIS, Silver Spring, MD.

Thorpe, R.B., J.M. Gregory, T.C. Johns, R.A. Wood, and J.F.B. Mitchell, 2001: Mechanisms Determining the Atlantic Thermohaline Circulation Response to Greenhouse Gas Forcing in a Non-Flux-Adjusted Coupled Climate Model. *J. Climate*, **14**, 3102-3116.

Vellinga, M., R.A. Wood, 2002: Global Climatic Impacts of a Collapse of the Atlantic Thermohaline Circulation. *Climatic Change*, **54**, 251-267.

Visbeck, M., H. Cullen, G. Krahmann, and N. Naik, 1998: An ocean model's response to North Atlantic Oscillation-like wind forcing. *Geophys. Res. Lett*., **25**, 4521-4524.

Wood, R.A., A.B. Keen, J.F.B Mitchell, an J. M. Gregory, 1999: Changing spatial structure of the thermohaline circulation in response to atmospheric CO_2 forcing in a climate model. *Nature*, **399**, 572-575.

Figure Captions

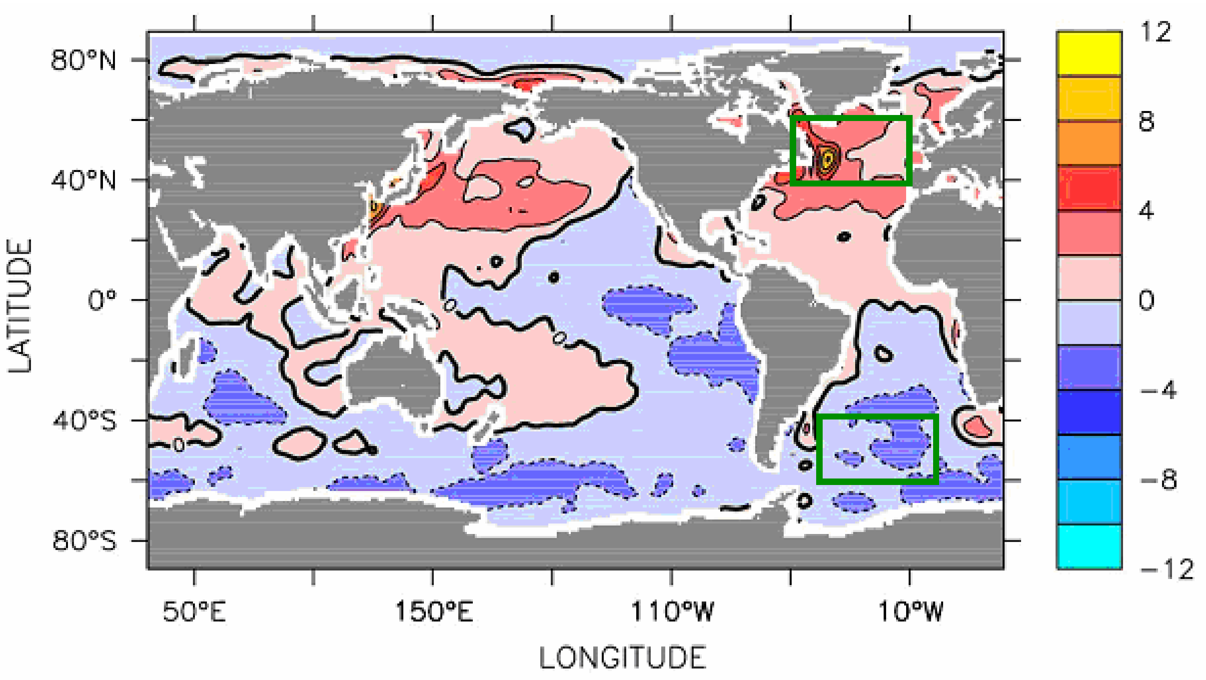
Figure 1: Linear trend in SST (°C/century) observed during the period 1980-2004. The global mean trend was removed to highlight the dynamical changes in the presence of global warming. A clear interhemispheric dipole is seen in the Atlantic which can be used as a fingerprint to detect changes in the MOC. The two boxes indicate the regions used to define the dipole SST anomaly index shown in Fig. 2.

Figure 2: Time series of the winter (DJFM) NAO index (shaded curve), a measure of the strength of the westerlies and heat fluxes over the North Atlantic and the Atlantic dipole SST anomaly index (°C, black curve), a measure of the strength of the MOC. The NAO index is smoothed with an 11-year running mean, the dipole index unsmoothed (thin line) and smoothed with a 11-year running mean filter (thick line). Multidecadal changes of the MOC as indicated by the dipole index lag those of the NAO by about a decade, supporting the notion that a significant fraction of the low-frequency variability of the MOC is driven by that of the NAO. The top figure shows annual data of LSW thickness (m), a measure of convection in the Labrador Sea, at ocean weathership Bravo, defined between isopycnals $\sigma_{1.5}$ = 34.72-34.62 following (Curry *et al*., 1998).

Figure 3: Cross correlation between the NAO index and the Atlantic dipole SST anomaly index as function of the time lag (years). The NAO index is based on winter values and the dipole index on annual values. No further time filtering was applied. The NAO leads the dipole index by about 5 to 20 years. Correlations above 0.2 are significant at the 95% level according to a t-test.

Figure 4: Sensitivity of the MOC (Sv) to changes in the potential density σ_0 (kg/m^3) of the Denmark Strait overflow obtained from simulations using two global ocean models (ORCA2 and ORCA05) with horizontal resolutions near 60° N of 120 (blue) and 30 km (black), and an Atlantic model (FLAME) with a horizontal resolution of 18 km (red). With each of these models, a sequence of experiments was performed addressing the response to imposed changes in the hydrographic conditions in the Nordic Seas. The data points depict the evolution of the overflow density at the sill, and the concomitant response of the maximum MOC transport (near 40° N) in the North Atlantic, after 15 (square symbol), 30 (diamonds), and 50 years (triangles) of integration, relative to reference states that correspond to MOCs of 16-18 Sv for the different model versions. According to the model simulations, the estimated decline of the overflow density based on hydrographic observations during the last 30 years (indicated by the gray bar) would correspond to a slackening of the MOC by about 1 Sv.

Figures



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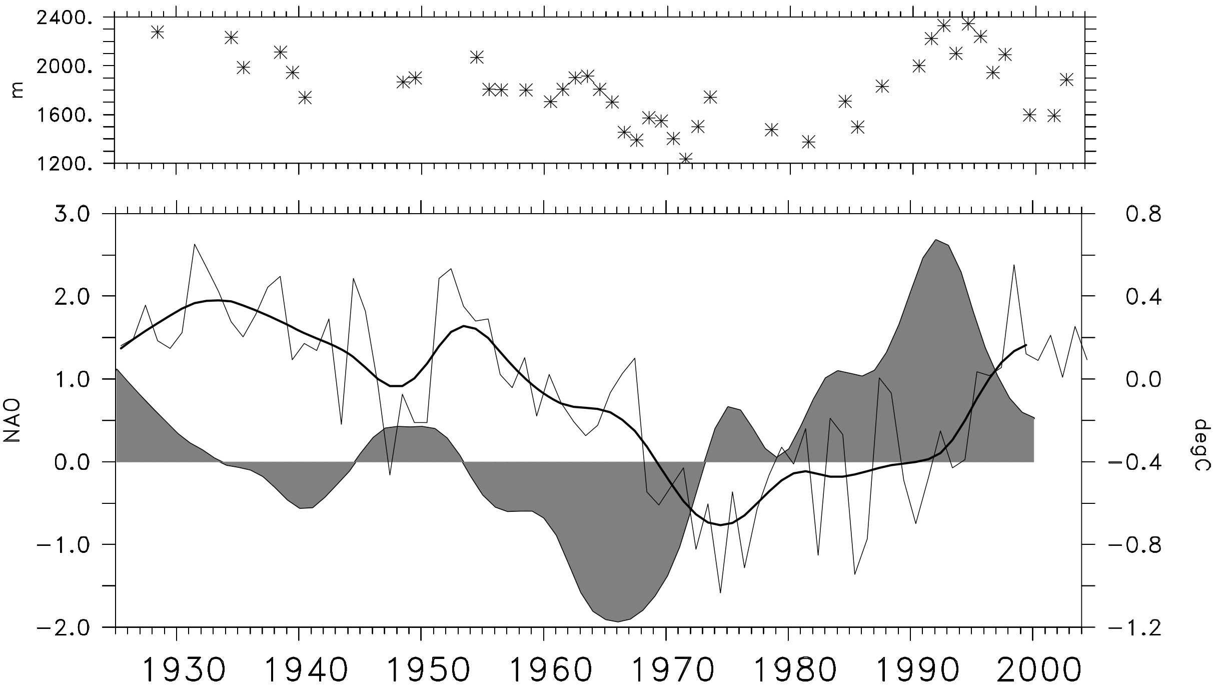

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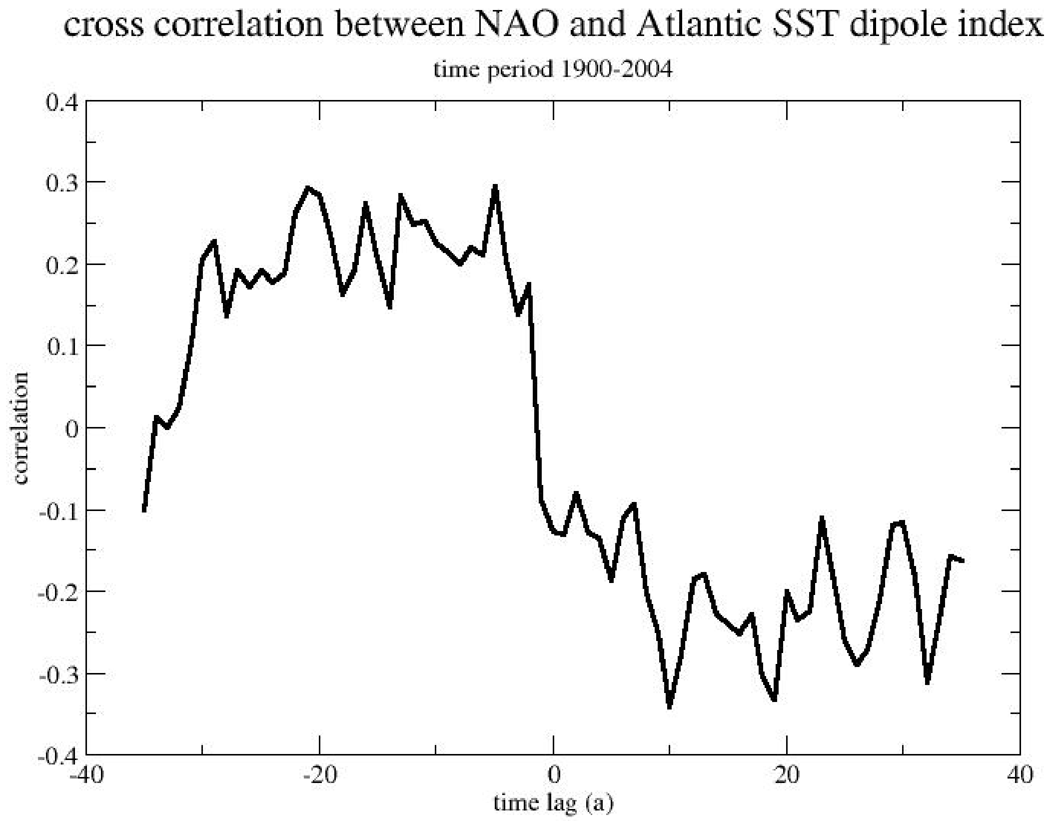

Figure 3: Cross correlation between the NAO index and the Atlantic dipole SST anomaly index as function of the time lag (years). The NAO index is based on winter values and the dipole index on annual values. No further time filtering was applied. The NAO leads the dipole index by about 5 to 20 years. Correlations above 0.2 are significant at the 95% level according to a t-test.

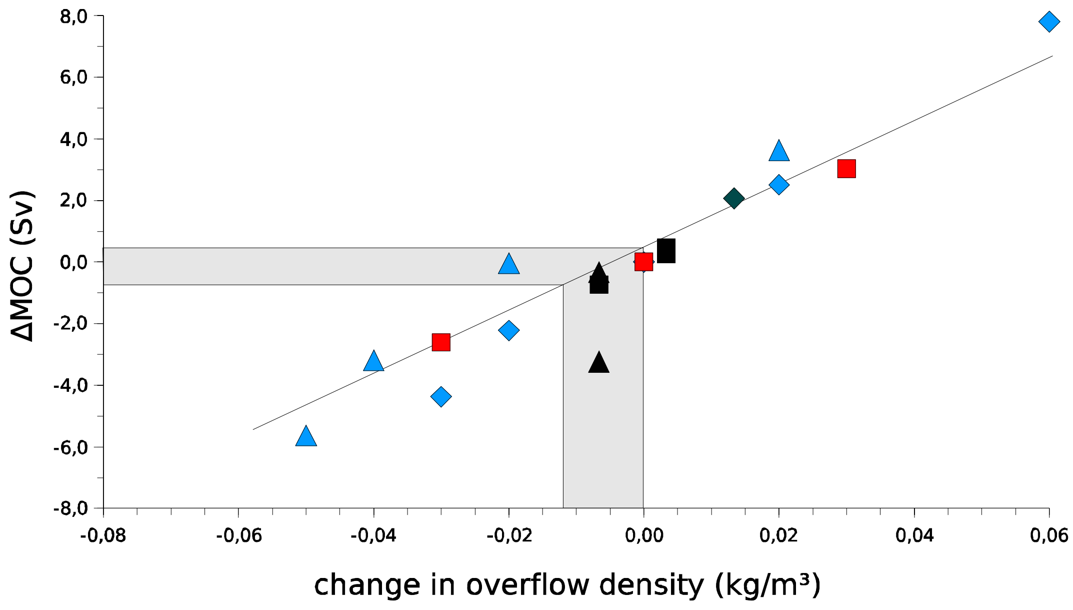

Figure 4: Sensitivity of the MOC (Sv) to changes in the potential density σ_0 (kg/m^3) of the Denmark Strait overflow obtained from simulations using two global ocean models (ORCA2 and ORCA05) with horizontal resolutions near 60° N of 120 (blue) and 30 km (black), and an Atlantic model (FLAME) with a horizontal resolution of 18 km (red). With each of these models, a sequence of experiments was performed addressing the response to imposed changes in the hydrographic conditions in the Nordic Seas. The data points depict the evolution of the overflow density at the sill, and the concomitant response of the maximum MOC transport (near 40° N) in the North Atlantic, after 15 (square symbol), 30 (diamonds), and 50 years (triangles) of integration, relative to reference states that correspond to MOCs of 16-18 Sv for the different model versions. According to the model simulations, the estimated decline of the overflow density based on hydrographic observations during the last 30 years (indicated by the gray bar) would correspond to a slackening of the MOC by about 1 Sv.