

## Oscillatory sensitivity of Atlantic overturning to high-latitude forcing

Lars Czeschel,<sup>1,2</sup> David P. Marshall,<sup>1</sup> and Helen L. Johnson<sup>3</sup>

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[1] The Atlantic Meridional Overturning Circulation (AMOC) carries warm upper waters into northern high-latitudes and returns cold deep waters southward. Under anthropogenic greenhouse gas forcing the AMOC is expected to weaken due to high-latitude warming and freshening. Here, we show that the sensitivity of the AMOC to an impulsive forcing at high latitudes is an oscillatory function of forcing lead time. This leads to the counter-intuitive result that a stronger AMOC can emerge as a result of, although some years after, anomalous warming at high latitudes. In our model study, there is no simple one-to-one correspondence between buoyancy forcing anomalies and AMOC variations, which retain memory of surface buoyancy fluxes in the subpolar gyre for 15–20 years. These results make it challenging to detect secular change from short observational time series.

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### 1. Introduction

[2] Warming of the upper North Atlantic over the past century has not been monotonic [Bindoff *et al.*, 2007]. Departures from steady warming on decadal time-scales make it mandatory to identify and understand modes of climate variability in order to anticipate and detect future anthropogenic climate changes. Multidecadal variability in North Atlantic sea surface temperature (SST) is well established [Kushnir, 1994] and is nowadays referred to as the Atlantic Multidecadal Oscillation. Model studies suggest that this variability is associated with changes in the AMOC [e.g., Delworth and Mann, 2000], with significant impacts on climate variability [e.g., Sutton and Hodson, 2005]. In this study we investigate the link between North Atlantic surface heat fluxes and the AMOC using an ocean circulation model and its adjoint.

### 2. Model

[3] The model used in this study is the MIT General Circulation Model [Marshall *et al.*, 1997] in a global configuration from 78°S to 74°N. The horizontal resolution is 1°, with 33 vertical layers ranging in thickness from 10 m at

the surface to 250 m at the bottom. It is comparable to the oceanic component of most state-of-the-art climate models. At the resolution employed here, ocean eddies are not explicitly resolved but parameterized following Gent and McWilliams [1990]. The model is driven by climatological monthly mean forcing obtained from NCEP/NCAR re-analysis [Kalnay *et al.*, 1996], i.e., the model has no externally forced interannual variability. Additionally, SST and sea surface salinity are relaxed to climatological values on a timescale of 30 days in order to prevent a large drift in watermass properties. A similar relaxation is also used in restoring zones at the boundaries in the Nordic Sea, Weddell Sea and the Strait of Gibraltar. SST patterns and ocean circulation features are broadly consistent with available observations.

[4] Conventional model sensitivity studies involve perturbing individual control variables (initial conditions, forcing, model parameters) so that, to assess the sensitivity to all control variables at all times, a huge number of experiments (here of order  $10^{12}$ ) is necessary. In contrast, the present model allows an adjoint calculation which gives the linear sensitivity of a cost function to all the control variables in a single integration, at all times between the time of the cost function evaluation and the time of the initial conditions [Marotzke *et al.*, 1999]. The adjoint is constructed by automatic differentiation [Giering and Kaminski, 2003; Heimbach *et al.*, 2005]; the cost function can be any scalar function of the model output, as long as it remains differentiable with respect to the control variables. The adjoint approach provides the sensitivity to small amplitude perturbations about a linearization of the underlying model, and hence we restrict our attention to modest forcing anomalies. The cost function used in this study is defined as the monthly-mean mid-latitude AMOC at 27°N (southward volume transport integrated between 1000 m depth and the sea floor). This has an annual mean of ~18 Sv and a mean annual cycle of ~5 Sv.

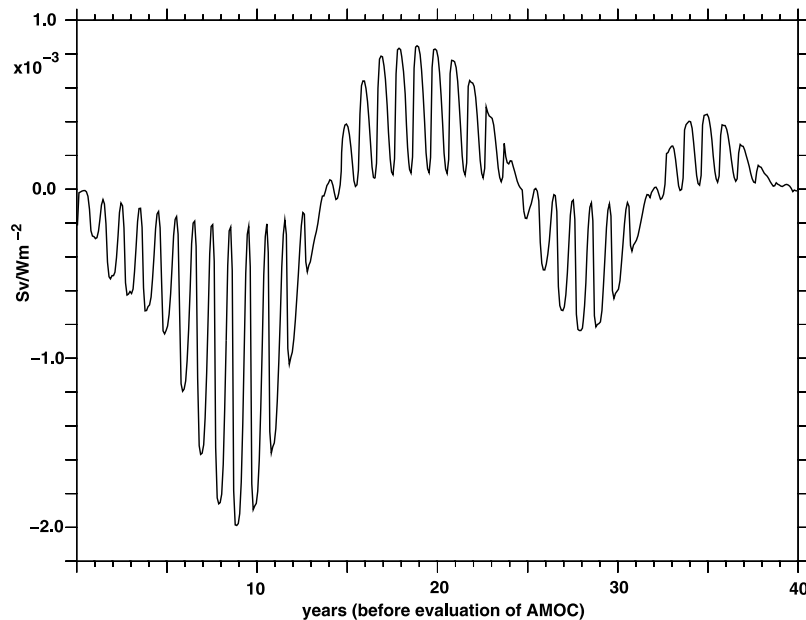
### 3. Results

[5] A prime mechanism for exciting a decadal oscillation in the AMOC and SST is heat flux variability over the North Atlantic, which is intimately connected with the North Atlantic Oscillation (NAO). Figure 1 shows the evolution of the sensitivity of the AMOC to the surface heat flux averaged over the subpolar gyre from 45°N to 70°N and 90°W to 10°E. The evolution is characterized by a damped decadal oscillation. In the 14 years before evaluation of the AMOC, increased heat loss over the subpolar gyre leads to an increased overturning. The strongest impact on the overturning variability results from the winter heat fluxes 9 years earlier. On the other hand, increased heat loss 15–25 years before evaluation of the AMOC leads to a decreased overturning. This result suggests

<sup>1</sup>Atmospheric, Oceanic and Planetary Physics, University of Oxford, Oxford, UK.

<sup>2</sup>Now at Leibniz Institute of Marine Science at Kiel University (IFM-GEOMAR), Kiel, Germany.

<sup>3</sup>Earth Sciences, University of Oxford, Oxford, UK.



**Figure 1.** Time series of the linear sensitivity of the AMOC at 27°N to surface heat fluxes averaged over the subpolar gyre. The sensitivities  $\partial\Psi_{27^\circ\text{N}}/\partial Q_{\text{net}}$  [ $\text{Sv/Wm}^{-2}$ ] are to a surface heat flux anomaly applied for one month and are integrated from 45°N to 70°N and 90°W to 10°E. A negative sensitivity means that cooling results in an increase in the AMOC. For example an additional cooling of  $1 \text{ Wm}^{-2}$  for one month over the given area at the minimum around 8.75 years would increase the overturning by  $2 \times 10^{-3} \text{ Sv}$ , 8.75 years later. The net induced AMOC variability is the integral over space and time of the sensitivities multiplied by heat flux anomalies. Note that linearity fails after 15–20 years (Figure 4).

that the heat flux induced variability of the AMOC at a given time is the combined response to the recent history of heat flux variability over the subpolar gyre.

[6] The integrated sensitivities are relevant for the case in which the subpolar gyre is forced with spatially homogenous heat flux anomalies. In order to calculate a more realistic heat flux induced AMOC variability we have superimposed inter-annually and spatially varying NCEP/NCAR heat flux anomalies on our sensitivities. This analysis suggests, for example, that the high NAO phase in the early 1990s led to an AMOC increase of about 0.3 Sv in 1999, but to an AMOC decrease of about 0.15 Sv in 2007, although the exact period and strength of the AMOC response is dependent on model parameters.

[7] A snapshot of the sensitivity of the AMOC at 27°N to temperature anomalies 8.25 years earlier, at a depth of 180 m, is shown in Figure 2. Negative values indicate that a cooler ocean would increase the AMOC 8.25 years later. The sensitivity pattern is complicated, and associated with a number of physical mechanisms operating in the forward model. For example, cold temperature anomalies in the subpolar ocean are advected from the south by the Gulf Stream system. Models of reduced complexity have shown that these anomalies induce, via thermal wind balance, a zonal overturning anomaly [TeRaa and Dijkstra, 2002]. The resulting downwelling anomaly at the eastern boundary causes a temperature anomaly of opposite sign. The anomalous zonal temperature gradient will then induce a positive AMOC anomaly which will transport anomalously warm waters at the surface towards the subpolar gyre, and the opposite phase of the oscillation will begin.

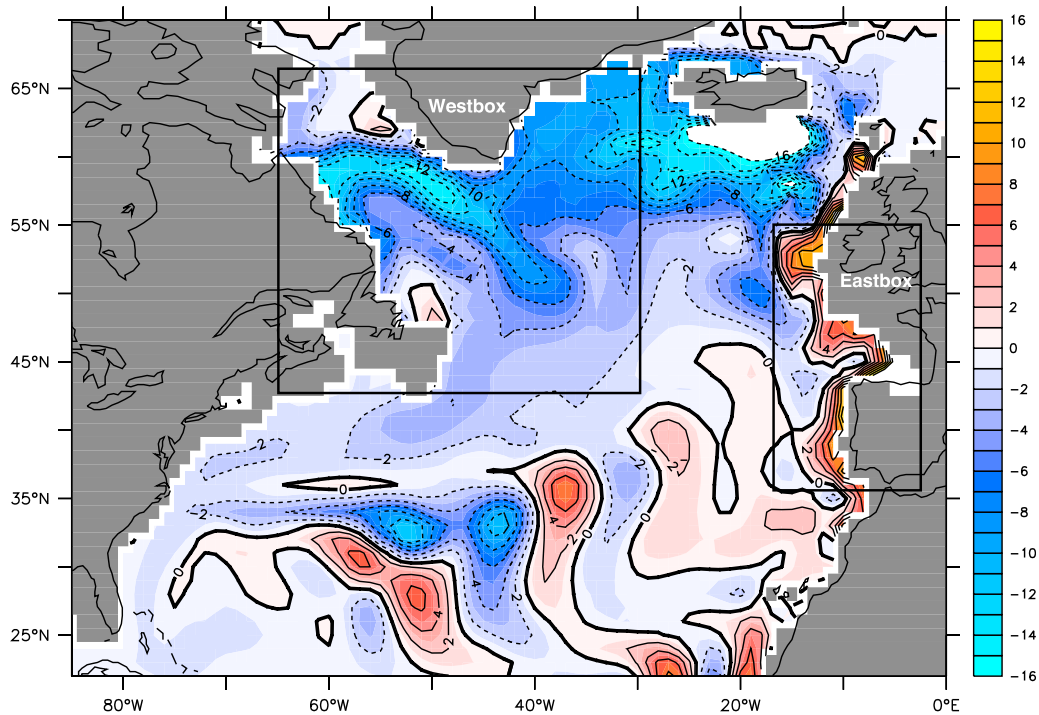
[8] However, temperature anomalies in the Eastbox also result in westward baroclinic Rossby wave propagation

along the southern flank of the Gulf Stream system. This manifests itself as a strong alternating signal in the sensitivities. These Rossby waves are prone to instability, and interaction with the mean current is likely. On reaching the western boundary, temperature anomalies are advected into the subpolar gyre by the Gulf Stream and play a role in changing the phase of the oscillation. Signal propagation due to Rossby waves along this path, and a continuation of the signal northward in the western boundary current, has also been identified in sea surface height data from satellite altimetry [Fu, 2004].

[9] The importance of Rossby waves is consistent with idealized models in which impulsive forcing produces a damped oscillatory response known as a Rossby-basin mode [Cessi and Louazel, 2001]. The timescale of such an oscillation, and its damping, is set by the longest Rossby wave basin-crossing time, which is about a decade. [Colin de Verdière and Huck, 1999] suggest that generalized ‘potential vorticity waves’, which rely on the mean stratification rather than the  $\beta$  effect, set the period. Faster boundary wave propagation may also play a role, propagating anomalies from high-latitudes to the equator in a matter of months [Johnson and Marshall, 2002]. Other studies suggest that advection may also result in an oscillatory response on decadal timescales [Eden and Greatbatch, 2003]. Animation S1 (of the auxiliary material) shows the evolution in time of the sensitivities shown in Figure 2.<sup>1</sup>

[10] The evolution of the sensitivity of the AMOC to temperature anomalies is shown in Figure 3. The sensitivity to anomalies in SST and temperature at 180 m has been

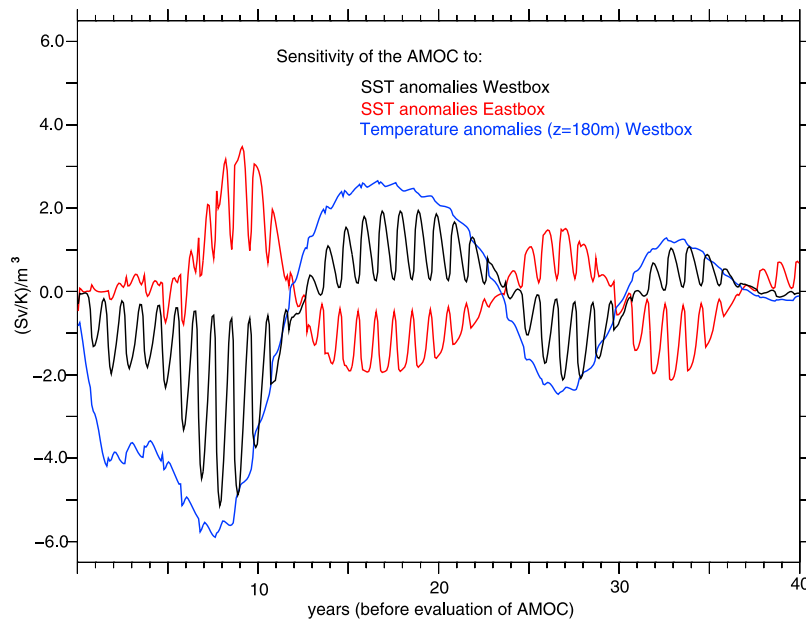
<sup>1</sup>Auxiliary materials are available in the HTML. doi:10.1029/2010GL043177.



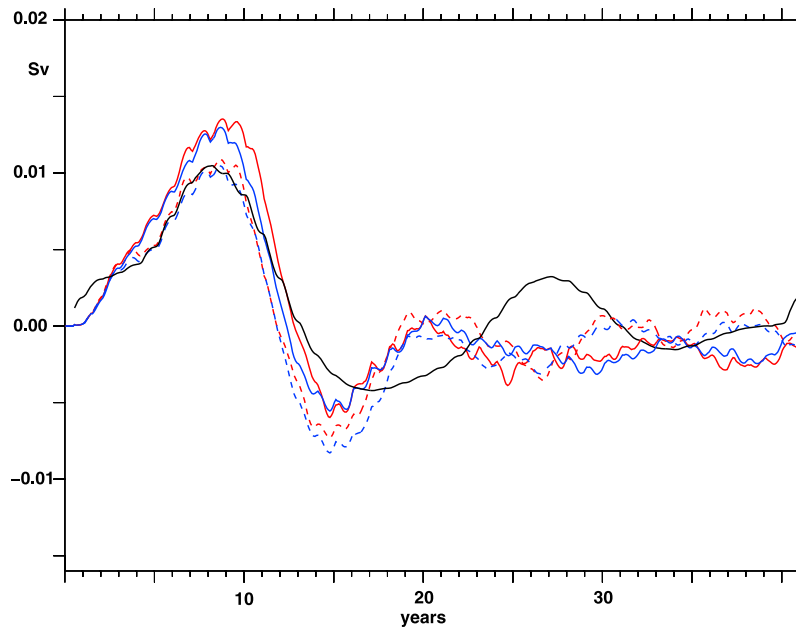
**Figure 2.** Linear sensitivity of the AMOC at 27°N to temperature anomalies introduced at 180 m depth, 8.25 years earlier. The sensitivity  $\partial\Psi_{27^\circ\text{N}}/\partial T$  is normalized to a temperature anomaly applied over a volume of 1 m<sup>3</sup> for one month. Units are [ $10^{-16}$  (Sv/K)/m<sup>3</sup>]. For example, a cold temperature anomaly of 1K applied for one year over a horizontal area of 10<sup>12</sup> m<sup>2</sup> in the Labrador Sea, and over a vertical extent of 10 m, results in a AMOC increase of roughly 0.1 Sv at 27°N, 8.25 years later.

averaged over the boxes shown in Figure 2. All three curves are characterized by the same damped decadal oscillation seen in the sensitivity to heat fluxes (Figure 1). The average sensitivity to SST in the Eastbox, which is the upwelling

region, is nearly the exact opposite of the sensitivity over the rest of the subpolar gyre, consistent with one of the physical mechanisms discussed above [TeRaa and Dijkstra, 2002]. The difference between the sensitivity to temperatures at the



**Figure 3.** Time series of the linear sensitivity of the AMOC at 27°N to sea surface temperature anomalies averaged over the Westbox (black) and Eastbox (red), and to temperature anomalies at 180 m over the Westbox (blue). All sensitivities are normalized relative to a temperature anomaly applied over a volume of 1 m<sup>3</sup> for one month. Units are [ $10^{-16}$  (Sv/K)/m<sup>3</sup>]. The Westbox and Eastbox are as shown in Figure 2.



**Figure 4.** Evolution of AMOC anomalies at 27°N in forward model integrations in which additional surface cooling (solid lines) or heating (dashed lines) of  $10 \text{ Wm}^{-2}$  (red) and  $15 \text{ Wm}^{-2}$  (blue) is applied in the Westbox. Shown is the difference in the AMOC [Sv] relative to the reference experiment. All results are normalized relative to a forcing anomaly which corresponds to an additional cooling of  $1 \text{ Wm}^{-2}$ . The black line shows the response expected from the linear adjoint model. The model response is linear when the curves are coincident. Whilst there is some early loss of linearity, the pattern of response is broadly consistent for the first 15–20 years. Thereafter, the response becomes unpredictable and the linear sensitivities are no longer appropriate.

surface and at 180 m is caused by restoring of SST to climatological data. Concomitant experiments made with a  $4^\circ$  resolution model (not shown), where the additional heat fluxes due to SST restoring were instead added along with the prescribed NCEP/NCAR heat fluxes, show higher sensitivities (up to 100%) compared with the same experiment with SST restoring, in agreement with Bugnion *et al.* [2006]. The experiment without SST restoring also shows a less pronounced seasonal cycle in the SST sensitivities. As the damping effect of the atmosphere is likely weaker than our SST restoring, reality may lie somewhere between these two experiments.

[11] The AMOC in our model is the integrated response to subpolar heat fluxes over multiple previous decades. In order to assess whether this response can be reconstructed from the linear sensitivities provided by the adjoint model, we run a suite of forward model integrations in which atmospheric heat flux anomalies of magnitude  $\pm 10$  and  $15 \text{ Wm}^{-2}$  are applied for 1 year (starting in summer, year 0) in the Westbox (Figure 2) which includes the area of deep convection. Figure 4 shows the resulting normalized anomalies in the AMOC at 27°N. While the model used here is neither time-translation invariant (due in part to the presence of a seasonal cycle in forcing) [Haine *et al.*, 2008], nor the sensitivity of the AMOC to subpolar gyre heat flux necessarily linear, an impulsive forcing does indeed generate an oscillatory response of similar magnitude and period to that predicted by the linear adjoint sensitivities. These forward model integrations suggest that the signal loses its memory of the forcing after about 15–20 years, broadly consistent with other recent studies (L. Zanna *et al.*, On the predictability and variability of the

Atlantic ocean due to optimal surface excitation, submitted to *Quarterly Journal of the Royal Meteorological Society*, 2010). However, the AMOC may still be affected by forcing on longer timescales, albeit in an unpredictable manner. Further experiments where the forcing anomalies are applied for longer time periods or in the Eastbox show similar results.

#### 4. Discussion

[12] Our study highlights that variability in the AMOC may retain memory of the forcing over the past 15–20 years. This memory has strong consequences for interpreting variability in the observed AMOC or in AMOC proxy data, such as the SST pattern over the North Atlantic. It is difficult, if not impossible, to attribute directly measured AMOC variability to heat flux anomalies which evolve in the subpolar gyre during one anomalous NAO phase. There is no simple one-to-one correspondence between AMOC variations and heat flux anomalies, and decadal variability in NAO-related heat (and freshwater) fluxes over the subpolar gyre will only result in AMOC variability of the same period if they project strongly onto the sensitivity function in Figure 1.

[13] Damped decadal oscillation of the AMOC sensitivity to high-latitude heat flux forcing is a robust feature in a whole series of sensitivity experiments performed with this model. It is the dominant pattern in sensitivity on longer time-scales for cost functions defined at a range of latitudes ( $48^\circ\text{N}$ ,  $27^\circ\text{N}$ ,  $8^\circ\text{S}$ ). We find no significant sensitivity to surface forcing outside the North Atlantic on time-scales of 5–40 years, consistent with previous forward model studies [Johnson and Marshall, 2004], but in contrast to P. Heimbach

et al. (Timescales and regions of the sensitivity of Atlantic meridional volume and heat transport magnitudes: Toward observing system design, submitted to *Deep Sea Research*, 2010) who find some sensitivity to interior temperature anomalies outside the North Atlantic. On shorter time-scales there is larger variability in the AMOC dominated by local wind stress anomalies [Köhl, 2005].

[14] In fact, the sensitivity to heat fluxes over the subpolar gyre is surprisingly small in our model. It is possible that atmospheric damping (i.e., SST restoring) is over-estimated in our model setup. While the oscillation is an internal oceanic mode, low frequency atmospheric forcing is crucial for its excitation [Delworth and Greatbatch, 2000]. Frankcombe et al. [2009] find spatial and temporal coherence in forcing anomalies is necessary for the amplitude of the variability to increase to realistic levels, broadly consistent with the spatial and temporal patterns of our adjoint sensitivities. A further reason for the small sensitivities could be linearization of nonlinear processes such as deep convection, although nonlinear forward integrations show sensitivities of the same magnitude (Figure 4).

[15] The oscillation described here also applies to salinity perturbations and is not restricted to near-surface buoyancy anomalies, with sensitivity to temperature at greater depths showing a similar oscillation. Thus, anomalies in overflows from the Nordic Seas are also likely to contribute.

[16] The period of the oscillation is depth dependent, suggesting different baroclinic modes and therefore different Rossby wave basin-crossing times; however, the largest buoyancy anomalies are generally found near the surface. In additional experiments (at coarser resolution) in which Rossby wave speeds are slower, the period of the oscillation in sensitivity is increased.

[17] Despite the robustness of the oscillation in our model experiments, the exact physical mechanism remains unclear and needs further investigation. We find evidence supporting the mechanism described by TeRaa and Dijkstra [2002], but also evidence for an important role of Rossby waves. Our different experiments show that both the period and damping timescale of the oscillation are dependent on model parameters, and determined by internal variability. The amount of damping is a crucial factor in estimating the relevance for the ocean and it will be interesting to determine whether eddy-resolving models show similar results. However, our findings are certainly important for the interpretation of contemporary climate models. For example, the results reported here support multidecadal predictability studies [e.g., Griffies and Bryan, 1997]. Adjoint sensitivity studies offer one route to unravelling and attributing the various contributions to North Atlantic climate change over the next few decades.

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L. Czeschel, Leibniz Institute of Marine Science at Kiel University (IFM-GEOMAR), Duesternbrooker Weg 20, D-24103 Kiel, Germany. (lczeschel@ifm-geomar.de)

H. L. Johnson, Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, UK. (Helen.Johnson@earth.ox.ac.uk)

D. P. Marshall, Atmospheric, Oceanic and Planetary Physics, University of Oxford, Parks Road, Oxford OX1 3PU, UK. (marshall@atm.ox.ac.uk)