Attributing variability in Atlantic meridional overturning to wind and buoyancy forcing

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An adjoint model is used to attribute variability in the Atlantic Meridional Overturning Circulation (AMOC) to wind and buoyancy forcing over the preceding 15 years. AMOC variability of magnitude ±5 Sv is excited by local wind forcing. In contrast, interannual to decadal AMOC variability of similar amplitude is excited by heat fluxes in the subpolar North Atlantic, with freshwater fluxes playing a very minor role. The magnitude of the reconstructed AMOC variability does not converge as the time window over which past forcing is accounted for is increased to 15 years. Beyond this point the assumption of linearity may break down, suggesting not all of the observed AMOC variability can be attributed in a linear manner to surface forcing. Nevertheless, the reconstructed AMOC variability is broadly consistent with RAPID observations at 26.5°N, especially in periods dominated by wind forcing.

1. Introduction

The Atlantic Meridional Overturning Circulation (AMOC), defined as the zonal-integral of the meridional volume flux, is believed to be a key driver of multidecadal variations in North Atlantic sea surface temperatures [e.g. Knight et al., 2005; Zhang and Delworth, 2006]. The AMOC has been continuously monitored at 26.5°N since 2004 and shows intense variability on all time scales [Cunningham et al., 2007; Kanzow et al., 2007]. A key outstanding challenge is to relate the observed variability to the past history of local and remote wind, heat and freshwater forcing.

The traditional forward modelling approach to attribution involves running ensembles of perturbed experiments, changing different aspects of the surface forcing in each case, and examining the impact on the modelled state. However, in a state-of-the-art ocean circulation model, the number of possible perturbations is vast, requiring a very large ensemble to perform a comprehensive sensitivity study. Beyond the technical difficulties of implementing a large ensemble investigation, attempts to identify causal mechanisms are often restricted to correlations that are only able to indicate mutual variability, without rigorous dynamical underpinning.

If we focus our attention on a specific metric of the evolved climate system, such as the AMOC at a particular latitude, it is helpful to also focus our computational resources on quantifying all potential causes of variability in this single objective function, as opposed to the impact of one imposed perturbation on the evolution of the entire climate state. For this, the adjoint method is a powerful tool [see Errico, 1997; Marotzke et al., 1999; Heimbach, 2008, for accessible formal descriptions], providing the local linear sensitivity of the objective function to all model inputs along the entire model trajectory. This gradient information thus reveals the linear transmission of influence through model space and time.

The adjoint approach has been employed by Köhl [2005], Czeschel et al. [2010] and Heimbach et al. [2011] to explore the sensitivity of the total northward AMOC transport across 27°N to buoyancy forcing. Czeschel et al. [2010] suggest that the AMOC may possess oscillatory sensitivity to high latitude heat fluxes on decadal timescales. Since sensitivities must be co-located in time and space with forcing anomalies to excite variability in the modelled transport, the implications of the long-term memory for the generation of AMOC variability has yet to be fully determined.

Recently Czeschel et al. [2012] have successfully applied the adjoint approach to elucidate the driving mechanisms of the seasonal cycle in the Florida Current transport. By multiplying the associated linear sensitivities by corresponding wind stress anomalies from reanalysis, the authors are able to model the current transport offline, recovering the observed phase and amplitude of seasonality. The memory of the modelled seasonal cycle is approximately three years; projecting increasingly historic forcing onto the linear sensitivities does not make a significant contribution to the transport estimate.

In this paper we follow a similar approach to attribute variations of the AMOC in an ocean circulation model to the past history of surface wind, buoyancy and freshwater forcing. The experimental approach is described in section 2. A brief review of the adjoint sensitivity pathways is presented in section 3. The AMOC is reconstructed by summing over the products of the adjoint sensitivities and climatological forcings, and compared with recent observations, in section 4. A brief concluding discussion is given in section 5.

2. Experiment Design

Our experiments are performed using a global configuration of the Massachusetts Institute of Technology general circulation model [MITgcm, Marshall et al., 1997] truncated below the poles. The horizontal resolution is 1° × 1°; the vertical is resolved by 33 levels of varying thickness increasing from 10 m at the surface to 250 m at depth. The model is driven by climatological monthly mean surface fluxes of heat, freshwater and momentum derived from the NCEP/NCAR Reanalysis II product [Kalnay et al., 1996]. The model is initialised to horizontally homogeneous temperature and salinity values based on hydrography. To prevent significant drift of sea surface temperature and salinity, we relax the simulated surface profiles to climatology...
with a damping timescale of 6 and 2 months respectively. Restoration of the full depth temperature and salinity fields towards climatology is also imposed at the open portion of the northern margin. Unresolved processes associated with mixing and advection by eddies are parameterised following Redi [1982] and Gent and McWilliams [1990] respectively. The K-Profile Parameterisation (KPP) scheme proposed by Large et al. [1994] is also employed to represent unresolved processes entailed in vertical mixing.

Following integration over 3000 model years, no significant trends are discernible in the tracer and momentum fields. The steady-state AMOC has an annual mean of approximately 23 Sv and a seasonal cycle with a peak-to-peak amplitude of approximately 7 Sv; the minimum occurs in February and the maximum in November. Seasonal variations north of approximately 20°N are principally described by a standing pattern of amplification. A decadal mode of AMOC variability is also visible that is weak relative to the seasonal cycle and is a common feature of both ocean-only and coupled-climate models, although the key driving mechanisms are disputed [e.g. Frankcombe et al., 2010].

We compute the objective function as the monthly mean northward volume flux across 25°N. Since there is a notable seasonal cycle in the equilibrated state, an ensemble of 12 adjoint calculations is required to probe the origins of monthly AMOC transport variability. The equilibrated model is integrated for a further 15 years and m months, at which point the objective function is analysed. By increasing the length of the forward integration by 1 month for each experiment in turn, the definition of the objective function is shifted from the January mean AMOC, \( \overline{\psi}_{25N}(m = 1) \), to the December mean AMOC, \( \overline{\psi}_{25N}(m = 12) \).

The adjoint model is obtained via algorithmic differentiation using the commercial tool TAF [Giering, 2010] and the gradient information is computed along the 15 years + m months trajectory. To obtain useful gradient information, we follow the standard practice of neglecting the highly nonlinear and discontinuous KPP scheme when forming the adjoint model [e.g. Hoteit et al., 2005] and also neglect the GM scheme for technical reasons.

To determine whether the linear sensitivities provide a meaningful description of perturbation growth in the nonlinear GCM, we compare the AMOC in an ensemble of forward integrations with perturbed surface forcing to that diagnosed offline using the linear sensitivities (not shown). Consistent with earlier experiments at a similar resolution [Czeschel et al., 2010; Heimbach et al., 2011], linear sensitivities are found to be representative up to lead times of approximately 15 years [also similar to the maximum growth timescales of surface-forced AMOC variability arising from nonnormal mode interaction, Zanna et al., 2012].

### 3. Sensitivity Pathways

The pathways of AMOC sensitivity to surface forcing have been previously discussed by Marotzke and Scott [1990], Bugnion et al. [2006a, b], Czeschel et al. [2010], and Heimbach et al. [2011]. Here we review the pertinent pathways to provide the context for the reconstruction of the AMOC in section 4. To simplify the discussion we present sensitivity distributions only for \( \overline{\psi}_{25N} \). It is natural to discuss sensitivity distributions with increasing lead time from the point at which the objective function is analysed. This reversal of the time axis allows adjoint waves to be examined moving away from the causal impulse [see Bugnion et al., 2006a]. As a result, propagation pathways observed in the forward model are traversed in the opposite direction in the adjoint framework. Following Heimbach et al. [2011] adjoint analogues of physics seen in the forward model are referred to as “dual” effect, and wind stress applied in the year prior to AMOC evaluation are shown in Figure 1. At a lead time of 1 month significant sensitivity to zonal wind stress is confined to the North Atlantic basin (A(i)) and dominated by a broad, zonally-uniform band of negative sensitivity extending across the basin width at 25°N. An increase in the westerly wind stress of 1 Nm−2 within this band would lead to a decrease in \( \overline{\psi}_{25N} \) by approximately 0.5 Sv in the same month by perturbing the Ekman transport across the basin. At a lead time of 3 months a signature of this fast barotropic response is still visible (A(ii)) although the sensitivity across 25°N is now positive and flanked by bands of alternating sign. Forward sensitivity experiments reveal that the physics responsible for this distribution is related to the steering of pressure perturbations around the Mid-Atlantic Ridge by barotropic Rossby waves [Pillar, 2013]. This mechanism is also responsible for the dipole straddling the Mid-Atlantic Ridge in the sensitivity to meridional wind stress (B(i)-(iii)).

At short lead times, notable sensitivities also occupy the upstream coastal waveguides (A(i), B(i)), highlighting the importance of trapped boundary waves in the rapid adjustment of the AMOC [e.g. Marshall and Johnson, 2013]. The sign of the sensitivity in the coastal waveguides can be understood by considering the orientation of the coastline (i.e. oneshore or offshore Ekman transport) and the contribution of the established pressure anomalies to the cross-basin pressure gradient at 25°N.

With increased lead time the dual boundary waves continue to propagate anti-cyclonically around the basin, passing through the equatorial waveguide and shedding dual coastal waves to the north and south (A(ii), B(ii)). These in turn shed eastward propagating dual Rossby waves into the ocean interior which show a notable equatorward tilt towards the eastern boundary due to the dual \( \beta \)-effect (A(iii), B(iii)). By a lead time of 1 year however, the amplitude of AMOC sensitivity to wind stress has diminished significantly (and does not re-amplify).

Snapshots of linear sensitivity of \( \overline{\psi}_{25N} \) to surface heat and freshwater fluxes are shown in Figure 1, columns C and D, and reveal similar rapid teleconnections to those discussed above (C(i), D(i)). A notable difference is the global extent of AMOC sensitivity to freshwater fluxes at a lead time of 1 month (D(ii)) due to the rapid redistribution of volume around the globe by barotropic waves [Lorbacher et al., 2012].

The modelled AMOC possesses multi-decadal memory to surface buoyancy forcing. The sign of sensitivity oscillates with forcing lead time in the North Atlantic; densification of the Gulf Stream and subpolar gyre at a lead time of 9 years strengthens the AMOC across 25°N (C(iii), D(iii)), but has an opposite effect at a lead time of 15 years (C(iii), D(iii)). This is consistent with the existence of a decadal mode of AMOC variability in the control (forward) integration. As discussed by Czeschel et al. [2010] the decadal variability here is due to a “thermal Rossby” mode [Huck et al., 1999; Te Raa and Dijkstra, 2002], modified by Gulf Stream advection and the passage of baroclinic Rossby waves radiated from the North-East Atlantic.

Interestingly, our results also suggest that buoyancy forcing over the Agulhas retroflexion - a key source region for the warm and salty waters forming the upper limb of the AMOC [Gordon et al., 1992] - is important at lead times nearing a decade and longer (C(ii)-(iii), D(ii)-(iii)). As suggested by Heimbach et al. [2011], the stationarity of the sensitivity here is possibly due to prolonged recirculation at a model resolution where leakage is not well represented.
4. Reconstructing the AMOC

The response of the AMOC to anomalies in atmospheric forcing, \( F \), may be computed from the adjoint sensitivities as follows

\[
\Delta \psi_{25N} = \int_0^{25 \text{ month}} \sum_{t=1}^{1 \text{ month}} \int \frac{\partial}{\partial F} \Delta \psi_{25N} \Delta F \, dy \, dx, \tag{1}
\]

where \( \partial \Delta \psi_{25N} / \partial F(x,y,t) \) are the monthly linear sensitivities and \( \Delta F(x,y,t) \) are monthly anomalies about the current NCEP/NCAR Reanalysis II climatological seasonal cycle (used to force the forward model) for the period 08/1983-08/2013 inclusive. By integrating over the global ocean and accumulating over lead time, \( t \), we obtain a timeseries of AMOC anomaly, \( \Delta \psi_{25N} \), due to each forcing (Figure 2) for the period 08/1998-08/2013 inclusive. By examining how the transport estimate changes as the maximum lead time, equivalently AMOC “memory”, is increased from 1 month to 15 years, we determine the importance of long-term memory in the generation of AMOC variability.

Significant sub-annual variability of the modelled AMOC is generated by anomalies in the surface wind stress (Figure 2a,b). The zonal and meridional wind stress generate transient fluctuations of approximately \( \pm 5 \) Sv and \( \pm 2 \) Sv respectively, on timescales less than a year. Between consecutive months these anomalies can differ by a magnitude of \( 6 \) Sv and \( 2.5 \) Sv respectively. The wind-driven variability is dominated by surface stress anomalies in the vicinity of \( 25^\circ \text{N} \) and the upstream coastal waveguides during the few months prior to AMOC evaluation. Since the AMOC adjusts rapidly to small amplitude surface stress variations (Figure 1A,B), the transport estimates are not significantly altered by accounting for increasingly historic wind forcing.

Lower frequency variability of the modelled AMOC is generated by anomalies in the surface buoyancy flux (Figure 2c,d). Due to the slow teleconnection pathways, the transport estimates are significantly altered by accounting for increasingly historic buoyancy forcing, although freshwater flux anomalies (about the seasonal climatology) appear relatively ineffective in stimulating transport variations [consistent with Hakkinen, 2002; Haak et al., 2003; Gregory et al., 2005]. In contrast to the potency of the instantaneous wind forcing, a buoyancy-forced AMOC anomaly exceeding \( 1 \) Sv is only excited after the response to surface heat fluxes (freshwater fluxes) is accumulated over 4 years (13 years) during the period examined. For the period 1998-2009 the buoyancy-driven AMOC anomaly is negative and amplifies as accumulated lead time is increased from 11 to 15 years (Figure 2c, orange to dark red lines). This weakening is driven principally by anomalous cooling over the subpolar gyre, associated with a predominantly positive phase of the NAO between the mid-1980s to 2000. The transport minimum in 2002-2003 is associated with strong surface cooling between 1987-1993 coinciding with the local minimum in subpolar sensitivity (Figure 1C(iii)).

The net linearised response of the AMOC for the period 08/1998-08/2013 (inclusive) exhibits monthly anomalies between -11 Sv and +6 Sv when accounting for all surface buoyancy and wind stress anomalies in the preceding 15 years (Figure 2e, dark red line). Annual mean anomalies for this timeseries range between -6.7 Sv and -0.4 Sv. It is important to note that the impact of heat flux anomalies applied at the maximum lead time considered in this study is detectable. This non-convergence of the transport estimate as the accumulated lead time is increased from 1 month to 15 years suggests we have not captured the full timescale over which the AMOC adjusts to surface heat forcing.

Whether our transport estimate is useful given the limitation discussed above, the reconstructed timeseries is compared to observations across the RAPID array, available for the period 04/2004-09/2012 inclusive (Figure 3). For a meaningful comparison to be made, the linearised response (Figure 2(iii)) has been superimposed on the background seasonal cycle in the forward integration. During periods when the surface heat flux makes a notable contribution to AMOC variability (e.g. 2004-2007), the reconstructed variability is weaker than that observed (Figure 3, black line) at 26.5°N. Furthermore, the mismatch between the observations and the reconstruction is more notable as the accumulated lead time is increased. However, during periods when the wind-forcing dominates AMOC variability, the reconstruction is broadly consistent with observed variability. In particular, the general weakening of the AMOC transport in 2009/2010 and the amplitude of the winter minimum in 2009/2010 and 2010/2011 (the so-called “double dip”) are comparable with published observations [McCarthy et al., 2012; Zhao and Johns, 2014] and primarily occur due to the anomalous zonal wind forcing in our analysis.

5. Concluding discussion

We have explored the atmospheric origins of subannual and lower frequency variability of the AMOC in a non-eddy ocean model using a numerical adjoint. We are able to explore the impact of forcing anomalies at lead times up to 15 years. Sensitivity distributions reveal the role of Rossby waves, coastally-trapped waves, and advective pathways in carrying disturbances generated at the ocean surface to \( 25^\circ \text{N} \). Consistent with previous studies, the modelled AMOC at this latitude possesses only short memory (of less than a year) to typical wind stress anomalies. In contrast, significant memory to buoyancy forcing persists on multidecadal timescales. The emergence of multi-annual timescales in the response is due to excitation of the thermal-Rossby type mode (section 3) by large-scale low frequency changes in the buoyancy forcing and may provide some limited predictability [Robson et al., 2012].

In considering the robustness of our results and directions for future study, it is important to note that the low frequency internal variability present in the control (forward) integration is dependent upon the model configuration, in particular the mixing coefficients and boundary conditions. This affects the simulated response to high-latitude buoyancy forcing (e.g. MacMartin et al., 2013) and consequently the details of the oscillatory sensitivity [Czeschel et al., 2010]. We suggest that this dependency may contribute to the increasing mismatch between the reconstructed and observed transport for extended accumulated lead time as seen in Figure 3 (for the period 2004-2007), although nonlinear interactions are also likely to play a role. Furthermore, eddy activity has recently been proposed as a driver of significant low frequency variability in the subtropical AMOC [Thomas and Zhai, 2013]. We are unable to address this issue at the present resolution.

We suggest that examination of the linearised response provides some useful lessons for attribution. In particular:

(i) the short memory of the AMOC to wind anomalies and the potency of local instantaneous wind forcing allows for easy attribution when the wind forcing is strong (e.g. in the case of the “double-dip” observed at the RAPID array).

(ii) The long memory (exceeding 15 years) of the AMOC to surface buoyancy forcing makes attribution a serious challenge during periods when the wind forcing does not
dominate AMOC variations, as illustrated by the non-convergence the reconstructed transport timeseries with increasing accumulated lead time (Figure 2(e), years 1998-2008).

(ii) Understanding observed AMOC variations may thus require knowledge of a long atmospheric forcing and oceanic response history. Due to lack of observations, the latter is only accessible using numerical models. However, the dependence of simulated low-frequency AMOC variability on model parameters and the importance of nonlinear interactions on long timescales complicates this task, as illustrated by the increased mismatch between modelled and reconstructed variability with increased accumulated lead time (Figure 3).

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Figure 1. Linear sensitivity of $\psi_{25N}$ to (A) zonal wind stress (B) meridional wind stress, (C) surface heat flux and (D) surface freshwater flux at the specified lead times (increasing from panel (i) to (iii) in each column). The colour scales are logarithmic. Absolute magnitudes smaller than (A,B) 0.0001 Sv/Nm$^{-2}$, (C) 1e-7 Sv/Wm$^{-2}$, (D) 100 Sv/ms$^{-1}$ have been set to 0. Positive sensitivity indicates that a strengthening of the (A) eastward wind, (B) northward wind, (C) upward heat flux, (D) upward freshwater flux at the specified lead time ($t < 0$) increases the overturning at 25°N.
Figure 2. Linearised response of the monthly mean AMOC, $\Delta \psi_{25N}$, to monthly anomalies in surface (a) zonal wind stress, (b) meridional wind stress, (c) heat flux, (d) freshwater flux and (e) buoyancy and momentum fluxes combined, computed from (1). Anomalies are computed as deviations from the Reanalysis II climatological seasonal cycle. Colour indicates the length of AMOC “memory” assumed when computing $\Delta \psi_{25N}$. Wind anomalies project onto sub-annual variability of the AMOC, whereas buoyancy anomalies excite lower frequency variability.
Figure 3. As for figure 2(e) except that the linearised AMOC response has been superimposed on the background seasonal cycle in the forward integration (annual mean transport removed). This allows a meaningful comparison with observed AMOC variability across the RAPID array (black line) for 04/2004-09/2012 inclusive (mean transport removed).