Influence of the Southern Annular Mode on projected weakening of the Atlantic Meridional Overturning Circulation

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ABSTRACT

Coupled climate models predict density-driven weakening of the Atlantic Meridional Overturning Circulation (AMOC) under greenhouse gas forcing, with considerable spread in the response between models. There is also a large spread in the predicted increase of the Southern Annular Mode (SAM) index across these models. Regression analysis across model space using eleven non-eddy-resolving models suggests that up to 35% of the inter-model spread in the AMOC response may be associated with uncertainty in the magnitude of the increase in the SAM. Models with a large, positive SAM index response generally display a smaller weakening of the AMOC under greenhouse gas forcing. The initial AMOC strength is also a major cause of inter-model spread in its response to climate change. The increase in the SAM acts to reduce the weakening of the AMOC over the next century by around 1/3, through increases in wind stress over the Southern Ocean, northward Ekman transport, and upwelling around Antarctica. The SAM response is also related to an increase in the northward salt flux across 30°S and to salinity anomalies in the high-latitude North Atlantic. These provide a positive feedback by further reinforcement of the AMOC. The results suggest that, compared with the real ocean where eddies oppose wind driven changes in Southern Ocean circulation, climate models underestimate the effects of anthropogenic climate change on the AMOC.
1. Introduction

The Atlantic Meridional Overturning Circulation (AMOC) consists of a northward flux of warm water in the Atlantic basin, which cools and sinks at high latitudes, returning southwards as dense water in the deep ocean (Wunsch 2002). Because it transports a large amount of heat northwards, it plays an important role in northern hemisphere climate (Vellinga and Wood 2002; Knight et al. 2005).

It is generally predicted that the AMOC will weaken in response to anthropogenic climate change (eg. Thorpe et al. 2001; Gregory et al. 2005; Cheng et al. 2013) with the potential for both regional and global climate impacts, such as moderation of global warming in Europe (Christensen et al. 2007; Meehl et al. 2007). Similar but larger changes in climate have been linked to the AMOC ‘bipolar seesaw’ during glacial periods (Broecker 1998).

AMOC strength, estimated using proxies such as $^{231}$Pa/$^{230}$Th and $^{14}$C (McManus et al. 2004; Robinson et al. 2005), is correlated with Arctic temperature as well as the intensity of Asian monsoons and climate over the Americas; and it is thought to be the driver of such changes, although modelling has proved inconclusive (Wang et al. 2001; Alley 2007; Seager and Battisti 2007; Broecker et al. 2010).

A weakening AMOC may also reduce the oceanic capacity for uptake of anthropogenic CO$_2$ via increases in North Atlantic stratification, and the associated weakening of the biological pump and decreased transport of CO$_2$ to depth (Schmittner 2005; Obata 2007; Zickfeld et al. 2008). The palaeoclimate record also hints that changes in AMOC strength are related to the capacity for terrestrial storage of methane and nitrous oxide, two potentially potent greenhouse gases (Flückiger et al. 2004; Sowers 2006; Wolff et al. 2010).

These physical and chemical factors make it important to understand how the AMOC is likely to vary in the near future; however, climate models differ significantly in the magnitude of their simulated AMOC and its response to climate change (fig 1). It is necessary to understand why this is the case if the global impacts of a changing AMOC are to be predicted with any skill.
Studies of the response of the AMOC to anthropogenic climate change have mainly been concerned with the effect of density fluxes and meridional steric height gradient in the North Atlantic (e.g. Mikolajewicz and Voss 2000; Thorpe et al. 2001; Gregory et al. 2005). However, numerous modelling studies have found that an increase in the strength of westerly winds over the Southern Ocean can result in an increase in the strength of the AMOC; this is through increased northward Ekman transport, Southern Ocean upwelling and deepening of the Atlantic pycnocline and/or via atmospheric processes (Toggweiler and Samuels 1995; Gnanadesikan 1999; de Boer and Nof 2005; de Boer et al. 2008; Delworth and Zeng 2008; Sijp and England 2009; Klinger and Cruz 2009; Marini et al. 2011; Wolfe and Cessi 2011). The relative contribution of the wind is uncertain (Kuhlbrodt et al. 2007), but it has the potential to affect the sensitivity of the AMOC to climate change.

A useful measure of the strength and location of these westerly winds is the Southern Annular Mode (SAM), the leading mode of atmospheric variability in the southern hemisphere (Gong and Wang 1999). It is described by the SAM index, often calculated from the difference in sea level pressure between subpolar and subtropical latitudes. Observational studies have shown a positive trend in the SAM index over the last few decades (Marshall 2003; Jones and Widmann 2004). Modelling studies also consistently predict a similar trend for the 21st century (fig 1), with a poleward shift and intensification of the southern hemisphere jet and a shift in storm tracks (Yin 2005; Fyfe and Saenko 2006). These trends are thought to be due to increases in atmospheric greenhouse gas concentrations and ozone depletion over Antarctica, which both act to increase temperature gradients in the upper troposphere and lower stratosphere, although the mechanisms are not fully understood (Shindell and Schmidt 2004; Toggweiler and Russell 2008; Son et al. 2009). There is some debate as to which effect dominates and which will have the more impact as ozone recovery progresses during this century; changes in ozone are thought to be particularly important during the austral summer and less important in the other seasons due to its strong seasonality (Marshall et al. 2004; Shindell and Schmidt 2004; Perlwitz et al. 2008; Son et al. 2008).
From a modelling perspective, these trends in the SAM might be expected to reinforce the AMOC via the mechanisms outlined above; they have the potential to reduce the impact of anthropogenic forcing on this important part of the climate system. However, these mechanisms are based on non-eddy-resolving models. The real Southern Ocean may be close to an eddy-saturated regime (e.g., Meredith and Hogg 2006), such that an increase in wind strength leads to an increase in southward eddy fluxes which compensate for the increased northward Ekman transport (Farneti and Delworth 2010). If the SAM is found to reinforce the AMOC in the climate models commonly used for prediction, it may be that the degree of AMOC weakening expected due to climate change is being underestimated.

To test these hypotheses, we carry out an investigation similar to Woollings et al. (2012) who found that the response of the northern hemisphere storm track to climate change could be related to the response of the AMOC. We invert their method, which consists of regressions across model space, to discover whether changes in the AMOC are related to changes in the southern hemisphere winds.

2. Methods

The discrepancy between different models’ predictions of climate is generally used to give an idea of the uncertainty in those predictions. The multi-model regression analysis used here is designed instead to take advantage of this spread in model results, by testing whether the models differ in a systematic way.

We use outputs from the model simulations run for the Coupled Model Intercomparison Project phase 3 (CMIP3), archived at the Program for Climate Model Diagnosis and Intercomparison (PCMDI). Out of the total 23 models we use up to 11 of the 12 that include the AMOC streamfunction in the readily available outputs (table 1). We exclude the iap_fgoals1_0_g model because its mean overturning circulation for the period 1960-99 is extremely weak and in the opposite direction to the other models. Each model has a different
number of ensemble members and therefore, for consistency, only the first ensemble member
of each model run is considered.

These models are not able to resolve mesoscale ocean eddies; this may affect the appli-
cability of the results to the real ocean because the Southern Ocean is thought to be eddy-
saturated (eg. Meredith and Hogg 2006; Munday et al. 2013). Nevertheless, at present,
models of this type are routinely used for comprehensive climate prediction, and it is there-
fore essential to understand their dynamics.

Annual mean data from the ‘20th Century Climate in Coupled Models’ (20C3M) scenario
was used to generate 1960-1999 mean values for a number of different variables (salinity,
potential temperature etc.) at each grid point (longitude, latitude, depth beneath ocean
surface). 2060-2099 means were found using outputs from the ‘Special Report on Emissions’
scenario SRESA1B. Initial conditions for this scenario are determined by the 20C3M results
and CO$_2$ is then increased approximately linearly to 720ppm by 2100, after which it is held
constant (Solomon et al. 2007).

We define a climate change ‘response’ in each variable at each grid point as the difference
between the 2060-2099 and 1960-1999 values. We do not extend the study beyond 2099
because not all models have outputs extending into the 22$^{nd}$ century. The forty year length
of each time period should ensure a representative value of the state of each variable.

All variables are re-gridded to a 2°latitude × 2°longitude grid to allow regression analyses
across model space at each grid point. Oceanic variables are also re-gridded to 33 specific
and unevenly spaced depth levels below the ocean surface. The re-gridding process should
not significantly influence the results, since we are looking at broad patterns and values,
rather than at individual grid points.

a. Climate Indices

Here we define indices that represent different modes of climate variability which will be
used as the independent variables in the regression analyses.
A simple SAM index is based on sea-level-pressure (SLP) (Gong and Wang 1999; Marshall 2003). It is the difference between the zonal mean SLP at 40°S and at 65°S:

\[ \text{SAM}_{\text{pressure}} = \overline{\text{SLP}}_{40^\circ S} - \overline{\text{SLP}}_{65^\circ S} \]

Based on figure 1 of Thompson and Wallace (2000) which demonstrates that changes in SLP over the Southern Ocean are well correlated with changes in wind strength, we define a second index:

\[ \text{SAM}_{\text{wind}} = \overline{U}_{45^\circ S-70^\circ S} - \overline{U}_{20^\circ S-45^\circ S} \]

where \( \overline{U}_{\text{Latitude}1-\text{Latitude}2} \) is the mean of the zonal wind velocity between latitudes 1 and 2. For both SAM indices we use the forty-year means (described above) of the austral summer months only: December, January and February (djf). In recent observational records and in models the tropospheric SAM signature has shown the largest trends in the austral summer period (Miller et al. 2006), which may be due to coupling between the troposphere and lower stratosphere (Thompson and Solomon 2002; Gillett and Thompson 2003). It is interesting to note that we find that the annual mean AMOC response is more highly correlated with the SAM\(_{\text{djf}}\) index response than with any other season. However, our results are robust if the regressions are carried out using the SAM index calculated for the whole year instead, since the response of the annual mean index is well correlated with the SAM\(_{\text{djf}}\) response in these models.

The response of the SAM\(_{\text{wind}}\) index is dominated by an increase in westerly wind strength between 45°S and 70°S, and is highly correlated with changes in the SAM\(_{\text{pressure}}\) index with \( R^2 = 0.92 \) and \( p = 2.8 \times 10^{-6} \). We use the SAM\(_{\text{wind}}\) index in the following regression analysis, and from here on simply refer to this as the SAM index.

It is thought that ozone depletion should enhance the positive trends in the SAM index, while ozone recovery during the 21\(^{st}\) century should oppose any increase in the SAM index (Perlwitz et al. 2008; Son et al. 2008). Comparing those models that include time-varying ozone with those that do not indicates that the SAM indices do not show consistently high
or low responses when time-varying stratospheric ozone is included. This may be due to
the particular set of models or choice of time periods used here. Either way, we conclude
that the presence or absence of time-varying stratospheric ozone is not a major factor in our
results.

The AMOC index we use is the maximum value of the annual mean AMOC streamfunc-
tion at 44°N, and below 400m to avoid surface maxima. This was chosen to be consistent
with the study of Woollings et al. (2012).

The North Atlantic Oscillation (NAO) is the dominant mode of climate variability in the
North Atlantic sector. A simple NAO index is the pressure difference between Iceland and
the Azores (Hurrell and Deser 2009).

\[ NAO = SLP_{Azores} - SLP_{Iceland} \]

To avoid unrepresentative, localised anomalies, area-mean rather than point values of SLP
are used (Iceland = 318 – 342°E, 54 – 60°N and Azores = 318 – 342°E, 26 – 36°N). As with
the SAM, we use the NAO\_djf index since the largest amplitude trends in SLP occur during
the boreal winter months (Miller et al. 2006; Hurrell and Deser 2009).

b. Regression

Every model displays a different response to anthropogenic forcing (2060–2099 minus
1960–1999) in its climate indices. At each grid point (longitude, latitude, depth), every
model has a slightly different response in any given climate variable. If these different climate
variable responses are plotted against the different responses of a climate index, a line of
best fit can be drawn. The accompanying statistics describe how closely the response of the
climate variable is related to the response of the climate index across model space. Here we
carry out the equivalent operation by calculating a linear regression of the response in any
given variable across model space on a vector comprising the normalised (to one standard
deviation) response of a climate index for each model; as in Woollings et al. (2012) and similar
to the methods of other authors (e.g. Hall and Qu 2006; Son et al. 2008). The gradient of the best fit regression line (B) is the change in the variable response per one standard deviation of the inter-model spread in the index response. The mean SAM response across the 11 models is 1.56 times larger than the standard deviation of this response; values of B should be scaled accordingly to allow comparisons of the magnitude of the SAM-related response of each variable with the multi-model mean response. A positive value of B suggests that models with the most positive index response also have the most positive response in the climate variable under consideration.

For some variables there are fewer than 11 models with available data (table 1). When this is the case the regression is carried out using the maximum number of models possible. Regressions are only carried out at grid points where every model used in that regression analysis has data.

The $R^2$ statistic is a measure of the proportion of the spread in the data that may be accounted for by the statistical model. $R^2$ is calculated for every regression. By the nature of the method, some grid points will have high $R^2$ values by chance. $R^2$ values are therefore only interpreted quantitatively when maps of $R^2$ show consistent values over a large area or when patterns of $R^2$ are similar to those of B.

Further statistical analyses are difficult due to the low number of models. The models and the climate indices are not strictly independent, so there is some uncertainty about the number of degrees of freedom compared to the number of models (Pennell and Reichler 2011). Estimates of the p-value are included, but these - as well as the estimates of explained variance from $R^2$ - may be overly confident. Values of B are often assumed to be statistically significant when $p < 0.05$ and white contours in maps of B (e.g. section 3) enclose areas where $p < 0.05$.

It is of course possible that regression relationships between two variables do not reflect a direct link but rather an indirect one in which both variables are related to some third factor. In particular, slight differences in model physics such as ocean diffusivity may lead to two
unrelated variables each having a particular response. However, since we generally regress
oceanic variables on atmospheric climate indices (i.e. spanning two different components
of the models), we consider it unlikely that the relationships we find are due to model
formulation in this way.

3. Results

a. AMOC-SAM relationship

The multi-model mean, 1960-1999 AMOC streamfunction (fig 2a) shows water flowing
north above 1100m, with southward flowing North Atlantic Deep Water (NADW) below.
The multi-model mean response of the AMOC to climate change (fig 2b) shows a maximum
weakening of the AMOC of around 6 Sv. The position of maximum weakening is deeper
than the maximum of the 20th century AMOC, suggesting that the models predict both a
weakening and shallowing of the AMOC.

Figure 2c shows the regression of the AMOC streamfunction response on the response of
the SAM. The coefficient B is positive almost everywhere showing that models with a large
positive SAM response generally display a smaller weakening of the AMOC under greenhouse
gas forcing. The scaled magnitude of the AMOC pattern associated with the SAM is opposite
in sign and around 1/2 of the mean AMOC response. If causal, this relationship suggests
that an increase in the SAM acts to significantly reinforce the AMOC under climate change,
reducing the maximum weakening by \( \sim 1/3 \). The significant \( R^2 \) values below 1000m (fig
2d) suggest that inter-model spread in the SAM index response can potentially explain up
to 35% of the inter-model spread in the mid-depth AMOC response, although this could be
an overestimate (see section 2.b).

We also regress the AMOC streamfunction response (across model space) onto a vector
comprising the 1960-99 mean AMOC index (fig 2e). The negative sign of the regression coeffi-
cient B shows that models with the greatest initial AMOC index show the greatest reduction
in the AMOC streamfunction, consistent with Gregory et al. (2005). The map of $R^2$ (fig 2f) shows that 40-75% of the inter-model spread in AMOC response in the deep-ocean is associated with the initial strength of the AMOC. Clearly, the different initial strengths of the AMOC in each model are important in determining its response to anthropogenic forcing, but cannot explain all of the spread. The relationship with the SAM has the potential to explain a large proportion of the remaining spread.

This reinforcement of the AMOC by the SAM is consistent with other non-eddy-resolving studies (eg. Gnanadesikan 1999; Delworth and Zeng 2008; Klinger and Cruz 2009; de Boer et al. 2010; Marini et al. 2011). These studies all find some form of causality for the relationship between southern hemisphere winds and the AMOC - although the mechanisms are not all the same - and we investigate the possibilities below. Our result adds to the evidence that wind forcing does indeed play an important role in the Atlantic overturning circulation, at least in coupled climate models. As far as we are aware, this is the first study to show this relationship in a multi-model analysis and under greenhouse gas forcing.

b. Physical Mechanisms

The studies given above suggest a variety of mechanisms via which the SAM may influence the AMOC. For simplicity, we separate these into three broad categories and investigate each one in the following sections.

Direct mechanical reinforcement of the AMOC by an increase in the SAM may be possible through increased northward Ekman transport out of the Southern Ocean and corresponding deepening of the Atlantic pycnocline (Gnanadesikan 1999; Klinger and Cruz 2009). This would be associated with increased upwelling around Antarctica, drawing water up from the depths of the deep ocean ridges where the NADW resides - due to restraints on geostrophic flow in mid-depths - and closing the circulation loop (Toggweiler and Samuels 1995).

It is also possible that an increase in the SAM might result in an increase in the wind-driven salt advection into the Atlantic Ocean and a subsequent strengthening of the AMOC.
through further northward advection of the salt (Marini et al. 2011; Sijp and England 2009).

The third potential linking mechanism is the influence of large-scale atmospheric telecon-
nections. The North Atlantic Oscillation (NAO) may have an effect on the strength of the
AMOC through changes in the wind-driven surface ocean circulation and surface buoyancy
fluxes at high latitudes (e.g. Dong and Sutton 2005; Mignot and Frankignoul 2005; Bellucci
and Richards 2006; Deshayes and Frankignoul 2008). The NAO and the SAM might be
expected to respond similarly to climate change, since temperature changes in the tropical
troposphere and in the stratosphere are believed to be important in both responses (although
more local temperature changes complicate matters). If they do, and if they both influence
the AMOC in the same way, it will be difficult to decipher which is the true driver. The
SAM has also been implicated in tropical precipitation processes linked to increasing Atlantic
salinity (Marini et al. 2011).

In order to establish which (if any) of these mechanisms are important, in the rest of
this section we consider the mean climate change responses of a range of variables, as well
as their regressions on the SAM response. We focus on the Southern and Atlantic oceans.

**Southern Ocean and South Atlantic mechanisms**

Since the SAM is the leading mode of climate variability in the southern hemisphere, one
might expect changes in the SAM index to be related to identifiable oceanic and atmospheric
responses here. Some of these responses - such as Ekman transport - are implicated in the
direct mechanical forcing mechanism described above.

The multi-model mean responses (2060-99 minus 1960-99) of six variables in the southern
hemisphere are shown in fig 3. Figure 4 shows B for the regressions of these six variables on
the SAM index. The maps of B show similar patterns and magnitudes (when scaled up by
1.56 - see section 2.b) to the mean response for every variable except salinity, suggesting that
the mean response in the southern hemisphere is well described by the SAM. The increase in
westerly wind strength between 40°S and 70°S (fig 3a) is the dominant factor in the increase
in the SAM index, with the increase in easterly wind strength between 20°S and 40°S playing a slightly smaller role.

The calculated Ekman transport regressed on the increasing SAM index (fig 4c) shows the expected enhanced northward flow out of the Southern Ocean into the closed Atlantic basin. $R^2$ values for the five remaining variables are somewhat inconsistent (not shown), but the patterns of $B$ are consistent with increased upwelling south of 55°S. Decreases in sea-surface-temperature (fig 4f) and increases in salinity (fig 4e) despite increased precipitation-minus-evaporation (P-E) (fig 4d) all support this conclusion. The barotropic streamfunction (fig 4b) suggests that there is also a spin up of the Antarctic Circumpolar Current (ACC) and of the southern hemisphere subtropical gyres, again expected due to the increasing SAM. The gyres also shift polewards, with the notable result of increasing the salinity and sea-surface-temperature in the Agulhas region of the southern Indian Ocean. These changes are all consistent with those found by other authors (eg. Hall and Visbeck 2002; Cai et al. 2005; Beal et al. 2011; Wang et al. 2011). The physically consistent response of each variable to an increased SAM suggests that parts of the mechanism neatly summarised by Gnanadesikan (1999) are indeed operating within these models. An increase in northward Ekman transport acts to generate upwelling in the Southern Ocean and an associated spin up of the ACC.

The increase in northward Ekman transport associated with the increase in the SAM index is around 1 Sv when summed across the southern boundary of the Atlantic basin. This is the same as the amount of water shown flowing northward above 2000m at 20°S in figure 2c, and suggests that much of the extra northward flow of the AMOC related to the SAM could be supplied by the Ekman transport due to increased Southern Ocean westerlies.

The magnitude of the AMOC response to a given wind forcing is consistent with values in the literature (eg. Gnanadesikan 1999; Johnson et al. 2007; Klinger and Cruz 2009). This, along with the findings above, lends weight to the idea that it is the direct mechanical forcing of the Southern Ocean by winds that is the reason for the SAM-AMOC relationship seen in the CMIP3 models.
If this is the case, there should be some response in the vertical density structure of the Atlantic that is related to the SAM. The multi-model-mean regions occupied by various water masses for the period 1960-1999 are evident from a plot of salinity (fig 5). The relatively saline water mass that fills much of the Atlantic basin is the North Atlantic Deep Water (NADW). The Antarctic Intermediate Water (AAIW) is the relatively fresh water mass which can be up to 2000m deep at 40°S and gets progressively shallower further north. The Antarctic Bottom Water (AABW) has a weak, deep and relatively fresh signature, although it is missing in several models.

The multi-model mean potential temperature (Θ) response (2060-99 minus 1960-99) shows an increase everywhere in the ocean (fig 6a). The salinity response (fig 6c) shows a predicted freshening of the NADW, due to the fresh surface water anomaly in the deep water formation regions. The northern hemisphere subtropical gyre becomes saltier, although the major increase in salinity is at shallow depths. This suggests a shallowing of its depth of influence and a reduction in downwelling in the gyre. The temperature response dominates over salinity in determining the density response in much of the ocean (fig 6e); however, the freshening of the high-latitude North Atlantic is important.

Cross-sections of the regression coefficient (B) for the same variables show that there is a cooling and freshening of a large mass of water between 2500m and 750m associated with an increase in the SAM (fig 6b and 6d). This is significant and associated with high $R^2$ values for both variables. It is dominated by changes near the deep western boundary. This supports the idea that an increase in the SAM is associated with a deepening of the lower branch (NADW) of the AMOC, as suggested by its regression on the SAM (fig 2c). A deepening of the relatively saline NADW should result in changing salinity transport, resulting in enhanced salinity in the new path of the NADW, and freshening above. It is unlikely that the change in properties is due to a consistent change in the location of deep water formation, because this is very different in each model. These plots suggest that oceanic properties such as salinity and temperature are responding to the SAM-driven
changes in the AMOC circulation, rather than forcing the circulation to change.

Despite the significant changes in $\Theta$ and salinity associated with the SAM, there is very little significant change in the associated density field (fig 6f). While this supports the conclusion above, it suggests that the pycnocline does not deepen in response to enhanced winds (as in Gnanadesikan (1999) and Klinger and Cruz (2009)). However, the vertical resolution in this study is too low to detect changes in pycnocline depth corresponding to a 2.5 Sv change in maximum AMOC strength.

All of our findings so far have been consistent with direct mechanical forcing of the AMOC by the SAM through the Southern Ocean. Below, we explore the remaining two categories of mechanism that could link the SAM and the AMOC.

**Northward salt flux across 30°S**

In order to test whether the salt flux into the Atlantic is affected by changes in the SAM, we calculate the change in the northward salt transport across 30°S into the Atlantic basin for each model:

$$\Delta \text{Salt Transport} = \Delta \sum_{\text{longitude}} \sum_{\text{depth}} (Sv),$$

where $S =$ salinity and $v =$ meridional velocity. The responses are then regressed on both the SAM index and the change in the maximum AMOC streamfunction at 24°S (fig 7). The regressions show a positive relationship between changes in the northward salinity flux and these two indices. For the regression on the SAM index, $R^2 = 0.66$ and $p = 0.0075$ and for the regression on the maximum AMOC streamfunction at 24°S $R^2 = 0.67$ and $p = 0.0071$. It seems that the salt flux increase is likely to be a feedback on the original AMOC strengthening generated by the increasing SAM index. Once the SAM has begun to reinforce the AMOC, the ocean acts to advect more salt into the Atlantic, further stabilising the overturning circulation. Changing properties of the Agulhas system (fig 4e and 4f) may also be involved, through increased salt leakage from the Indian Ocean into the Atlantic,
although the effects of this are extremely difficult to quantify using our methodology.

**North Atlantic mechanisms**

Figure 8 shows the multi-model mean response (2060-99 minus 1960-99) in the North Atlantic for the same six variables that were plotted in the southern hemisphere in figure 4. Zonal surface wind ($U_{surf}$) displays a north-south dipolar pattern (fig 8a), (also apparent in the Pacific) which has a similar pattern to the wind anomalies associated with a positive phase of the North Atlantic Oscillation (NAO) (Hurrell and Deser 2009). The barotropic streamfunction response (fig 8b) indicates a northward shift and/or weakening of the subtropical gyre in the North Atlantic, potentially associated with the changes in the wind in figure 8a. This is consistent with the reduction of downwelling in the subtropical gyre suggested by figure 6c. The subpolar gyre response is much less clear. Precipitation minus evaporation (P-E) also shows the NAO-like dipolar pattern over both the Atlantic (fig 8d) and Pacific and is therefore likely to be a result of intrinsic atmospheric processes. A broad freshening occurs at 100m depth in the area south of Greenland (fig 8e), presumably caused by the increase in P-E there (eg. Manabe et al. 1991). A tongue of fresh water that extends south along the African coast appears to be the result of the northern freshening being advected southwards by the subtropical gyre. Sea-surface-temperature indicates a predicted warming of almost the entire ocean (fig 8f). The exception is in the high-latitude North Atlantic where the warming is hampered due to the reduction of the AMOC circulation and its associated heat transport; this is a common result (Meehl et al. 2007; Drijfhout et al. 2012). The changes in sea-surface-temperature and salinity in the deep water formation regions are thought to be the cause of the predicted reduction in AMOC strength (Dixon et al. 1999; Gregory et al. 2005).

Maps of the coefficient (B) for each variable regressed on the SAM index show different features to the mean responses (fig 9). Neither $U_{surf}$ (fig 9a), Ekman transport (fig 9c) nor the barotropic streamfunction (fig 9b) show any clear pattern, and there is no significance
at the 95% level for the majority of grid points. However, salinity shows a well defined and
significant pattern (fig 9e); a positive anomaly exists in the position of the subpolar gyre
in the models, with a negative anomaly in the subtropics. The North Atlantic is the only
region of the northern hemisphere where such high significance is seen, suggesting that the
results here are indeed important. High latitude increases in salinity (fig 9e) and density
(not shown) associated with an increase in the SAM are consistent with an increased AMOC.
However, this relationship with the AMOC could be cause (since increasing density favours
deep water formation) or effect (since increasing the AMOC should lead to an increase in
northward salt advection).

We address this issue by looking at the maps of B for sea-surface-temperature and for P-E
(figs 9d and 9f). The patterns of these two variables and salinity are qualitatively similar,
with a region of increased sea-surface-temperature, reduced P-E and increased salinity over
the subpolar gyre, and a two-lobed pattern (where $\Theta_{\text{surf}}$ and salinity have opposite signs)
extending south along the eastern boundary of the ocean. The simplest explanation for
the similarity of these patterns is that the changes in sea-surface-temperature associated
with the SAM are due to advection of warmer water from the south, which is also relatively
salty. The temperature anomalies drive anomalies in P-E, which in turn reinforce the salinity
anomalies.

While it is possible that these changes in ocean surface temperature are caused by at-
mospheric anomalies in the northern hemisphere linked to the SAM, this seems unlikely
considering the lack of consistent pattern in B for $U_{\text{surf}}$ regressed on the SAM response (fig
9a). It is more likely that the temperature changes are a result of increased advection of
warm water into the subpolar gyre from the south due to the reinforcement of the AMOC. If
this is the case then the salinity anomalies in the high latitude North Atlantic related to the
increase in the SAM are another feedback process - like the salt advection at 30°S - which
can further strengthen the AMOC, rather than a primary linking mechanism.

The lack of a clear North Atlantic surface wind response to the SAM also suggests that
the SAM and NAO responses across the 11 models are unrelated. Values of $R^2 = 0.026$ and $p = 0.63$ for a regression of the NAO index response on the SAM index response show that this is indeed the case. The two indices do not evolve in a similar way in this set of models, and the SAM is much more strongly related to changes in the North Atlantic ocean than in the North Atlantic atmosphere, suggesting that the latter plays little role in the forcing of the AMOC by the SAM. Moreover, regressions of the AMOC response on the NAO index (not shown) are nowhere significant, suggesting that the NAO does not play as large a role as the SAM in modulating the strength of the AMOC under climate change in this set of models. Similarly, neither figure 4 nor figure 9 show significant links between the response of tropical variables and the SAM response, suggesting that the tropics do not link the SAM and the AMOC here, although other authors have found links between the SAM and tropical processes (eg. Thompson and Lorenz 2004).

The absence of any clear inter-hemispheric, atmospheric response to the SAM suggests that the relationship between the SAM and the AMOC responses to greenhouse forcing stems primarily from the direct mechanical forcing of the AMOC by the SAM via Southern Ocean processes.

4. Discussion

a. Note on the model subset and its relation to a wider range of models

To assess whether our results are reproducible using more of the CMIP3 models, we have carried out a regression analysis of sea-surface-temperature for the 19 models where this data is available. The global pattern of the coefficient (B) for the sea-surface-temperature response regressed on the SAM index response is almost identical to that when using only 11 models (not shown). SAM-related cooling is observed around Antarctica, with increasing temperature in the Agulhas system and North Atlantic (the North Pacific shows a cooling). Values of p and $R^2$ are lower in the North Atlantic for this regression, but we argue that
the similarity of the global pattern demonstrates that our results would be similar had more models been used in the analyses. It is also important to note that individual models used for the CMIP3 have been used in other studies (using alternative methods) where the SAM has been found to reinforce the AMOC (Delworth and Zeng 2008; Marini et al. 2011). Our study complements and extends these findings to a wider range of CMIP3 models.

We also consider the relevance of these results to the newer and larger set of CMIP5 models, for which publications are beginning to emerge. They typically have better resolution than the CMIP3 models (Taylor et al. 2012), and all contain time-varying stratospheric ozone (Gillett and Fyfe 2013). However, their ocean components are still not eddy-permitting, and despite the better representation of ozone changes, the SAM index still exhibits an increase in all seasons as the atmospheric greenhouse gas concentration increases (Gillett and Fyfe 2013). Like the CMIP3 models, the CMIP5 models all predict weakening of the AMOC (Cheng et al. 2013), and there is a poleward shift of the southern hemisphere subtropical gyres linked to the increasing SAM index (Meijers et al. 2012). These features of the CMIP5 model data suggest that they behave in very similar ways to the CMIP3 models, with some improvements with regards to ozone. We suggest that if the same analysis were to be carried out on the CMIP5 models, the results would be similar to those obtained here.

b. Drivers of spread in the SAM response

In order to better understand the full set of processes operating here, it is useful to consider what may drive the spread in the SAM response between models.

Firstly, it may be the case that models with greater sensitivity to greenhouse gas forcing display a greater response in the SAM index. The relationship between the SAM index response and the transient and equilibrium climate sensitivities is positive as might be expected, but neither is significant (we find p-values of $p = 0.27$ and $p = 0.30$ for regressions of the SAM index response on the transient and equilibrium climate sensitivities, respectively). This shows that differences in model SAM responses cannot be sufficiently explained by
differences in climate sensitivity. Normalising the responses of the climate variables by the
global mean temperature change also has no significant impact on the results.

Secondly, we have found that in this set of models the SAM response is not obviously
related to whether the models do or do not contain time-varying stratospheric ozone. Green-
house gas emissions appear to be the dominant driver of changes in the SAM. However, this
is inconsistent with other studies that have found that changes in ozone play an important
role in observed and modelled changes in the SAM (Shindell and Schmidt 2004; Miller et al.
2006; Perlwitz et al. 2008; Son et al. 2009). This discrepancy is likely to be due to the exact
subset of models used, and the choice of time periods compared here.

It has been shown using a different subset of the CMIP3 climate models that the response
of the SAM is related to its initial state (Kidston and Gerber 2010), although this was
less clear in the austral summer months due to differences in model representation of the
stratosphere. We carry out a similar test and find that, if the anomalous giss-aom model is
excluded, models with the weakest initial SAM show the biggest response, consistent with
the above study. (The giss-aom model has a very low initial SAM index, and also displays the
smallest response to climate forcing of any of the models, falling well outside the relationship
defined by the other models.) This relationship is significant \( p = 0.035 \) even though we
are using the austral summer months. This suggests that in general, the initial state of the
SAM does indeed play a role in determining its response.

c. Eddy activity in the Southern Ocean

Because of the coarseness of the resolution of these CMIP3 climate models, they do not ex-
plicitly represent ocean eddies. Eddies are either parameterised using different versions of the
Gent-McWilliams (GM) parameterisation (Gent and McWilliams 1990; Visbeck et al. 1997)
or are not represented at all. Of the set of models used here, cgm3.1(t47), cgm3.1(t63),
csir-mk3.0, miroc-3.2(medres), echam/mpi-om and mri-cgcm2.3.2 use constant GM coeffi-
cients, gfdl-cm2.1 and ipsl-cm4 use varying but capped GM coefficients, bccr-bcm2.0 is an
isopycnal model that uses interface smoothing equivalent to a GM parameterisation, and giss-aom and inm-cm3.0 do not contain eddy parameterisation (Kuhlbrodt et al. 2012).

This presents a challenge for relating these results to the real ocean. Eddies extract energy from sloping density surfaces and act to flatten them (Lee et al. 1997). This means that in the Southern Ocean there is a time-varying bolus circulation that opposes the northward Ekman transport due to westerly winds. This has two implications for the climate change response of the Southern Ocean.

Firstly, eddies may respond to changes in wind forcing by opposing changes in isopycnal slope, making the transport of the ACC relatively insensitive to wind forcing (e.g. Hallberg and Gnanadesikan 2006). Based on observational and modelling studies, it has been suggested that although the Antarctic Circumpolar Current (ACC) responds to changes in wind stress on short time scales (days to years) (Hughes et al. 2003; Meredith et al. 2004), it varies by a proportionally much smaller amount over longer time-scales (Meredith et al. 2004; Meredith and Hogg 2006; Hogg et al. 2008; Böning et al. 2008; Munday et al. 2013). Such a phenomenon is known as eddy-saturation.

Secondly, in an eddy-permitting model, the bolus circulation has been shown to increase as the strength of the westerly winds increases (Farneti and Delworth 2010). This counteracts the increased northward Ekman flow and therefore greatly reduces the effect of an increase in the westerly winds on the AMOC (eddy-compensation).

Better agreement between models with eddy parameterisation and those that are eddy-permitting can be achieved if the GM coefficient is allowed to vary and the upper limit on the coefficient is removed (Farneti and Gent 2011; Gent and Danabasoglu 2011). However, in such models the GM coefficient is generally dependent on isopycnal slope and therefore, while the models may be eddy-compensated, this requires a change in the isopycnal slope such that they cannot also be eddy-saturated. The mean climate state of the models can be adversely affected (Farneti and Gent 2011). In any case, none of the models in this study have uncapped GM coefficients and are therefore unlikely to capture the full extent of eddy
behaviour under wind stress forcing.

There seems to be a growing consensus that the Southern Ocean is close to eddy-saturated. If this is the case then the effect of an increasing SAM on the ACC and on the AMOC in the 21st century may be rather limited. Although eddy compensation may not be so closely tied to eddy saturation as many authors assume (Meredith et al. 2012), its likely underestimation in these models has implications for their skill in predicting the AMOC response of the real ocean.

d. Are the predictions of the AMOC in climate models too conservative?

In the previous two sections we have outlined two processes that are not well or consistently represented in the CMIP3 models used for climate prediction; ozone recovery and ocean eddy activity. Both of these processes have a possible bearing on our results.

Firstly, it has been shown in general that models with ozone recovery display weaker trends in the SAM in the 21st century (Miller et al. 2006; Perlwitz et al. 2008; Son et al. 2009), yet not all CMIP3 models include this. This may mean that the observed increasing trend in the SAM will not continue at such a rate if the effect of greenhouse gases is counteracted by ozone recovery in the future. If the SAM-AMOC relationship found here is representative of the wider range of models, this would mean that the CMIP3 climate models may be underestimating the full magnitude of AMOC decline. This hypothesis could be directly tested by analysing the CMIP5 models, which all include time-varying ozone as discussed above.

Underestimation of AMOC weakening may be compounded by the effect of eddies in the real ocean, which may reduce the influence of the SAM on the AMOC through eddy compensation. If the magnitudes of the regression results obtained here are to be believed, the models may be underestimating the AMOC reduction by up to 1/3. This could have important implications for predicted ocean heat transport and both European and global climate, as well as potential implications for carbon cycling between the ocean and atmosphere. Note
that the coarse resolution of the models may mean that other drivers of AMOC change are also misrepresented. Future studies should aim to better quantify the potential for, and consequences of, underestimation of the AMOC weakening in CMIP5 compared with higher resolution models.

5. Summary

When subjected to greenhouse gas forcing, the Southern Annular Mode (SAM) response of 11 climate models has been shown to reinforce the Atlantic Meridional Overturning Circulation (AMOC) through the mechanisms of increased northward Ekman transport and upwelling in the Southern Ocean. These processes force a more vigorous AMOC, as well as increasing the salt flux across 30°S into the Atlantic. Once the AMOC has begun to respond to the southern hemisphere westerlies, a system of feedback processes is set in motion in the North Atlantic, involving warming of northern waters and associated evaporation and salinification. These processes increase the density of waters near the regions of deep water formation, favouring deep convection and further AMOC reinforcement.

The overall effect of the increasing trend in the SAM is to reduce the predicted weakening of the AMOC by around 1/3. The inter-model spread in the SAM response can potentially explain up to 35% of the spread in AMOC responses, although this could be an overestimate. A large proportion of the spread in the response of the AMOC can also be explained by differences in its initial state. The same is true of the response in the SAM.

The effect of ocean eddies is likely under-represented in the models investigated here, and this may mean that predictions of the weakening of the AMOC in these models are too conservative by up to 1/3. The lack of ozone recovery in some CMIP3 models may contribute to this effect. Further work is required on the nature of eddy compensation in the Southern Ocean to assess the skill of coarse resolution models and to investigate whether they underestimate the weakening of the AMOC or whether the SAM may indeed reduce
the effects of climate change in the North Atlantic.

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REFERENCES


Hughes, C. W., P. L. Woodworth, M. P. Meredith, V. Stepanov, T. Whitworth, and A. R.


Selected model features of the eleven models used in the analysis; including the original ocean and atmosphere resolution and equilibrium and transient climate sensitivities. We have also included a list of the parameters missing from the available outputs of each model. The identifiers in the final column indicate whether the model includes (Y) or does not include (N) time-varying stratospheric ozone in the 21st century. Resolutions and sensitivities were taken from the 2007 IPCC WG1 report (Solomon et al. 2007), and details of ozone prescription from Son et al. (2008) and Son et al. (2009).
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<table>
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<tr>
<th>Model</th>
<th>Ocean Resolution ((^\circ\text{Lat} \times ^\circ\text{Long}))</th>
<th>Atmosphere Resolution ((^\circ\text{Lat} \times ^\circ\text{Long}))</th>
<th>Equilibrium Climate Sensitivity ((^\circ\text{C}))</th>
<th>Transient Climate Sensitivity ((^\circ\text{C}))</th>
<th>Missing Parameters</th>
<th>Time-varying Ozone</th>
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<td>2.2</td>
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<td>3.4</td>
<td>1.5</td>
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<td>3.2</td>
<td>2.2</td>
<td>potential density</td>
<td>N</td>
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</table>
1 Modelled evolution of the a) maximum Atlantic Meridional Overturning Circulation (AMOC) streamfunction at 44°N and deeper than 400m and b) the fifteen-year moving-average of the Southern Annular Mode (SAM) index, defined here as the $\text{SAM}_{\text{wind}}$ index given in section 2a. The large spread in SAM index values is due to the different latitudinal extent of the band of westerlies vs the band of easterlies in the southern hemisphere. The time series are plotted from 1871 to 2100 for all 12 models considered in this study. The name of each model is indicated in the legend. The 20th Century Climate in Coupled Models (20C3M) simulations are used for 1850-1999 and the SRESA1B projections after this.

2 a) Multi-model mean Atlantic Meridional Overturning Circulation (AMOC) streamfunction (STF) for the period 1960-1999. b) Multi-model mean AMOC streamfunction response from the period $(1960-1999)_{\text{mean}}$ to the period $(2060-2099)_{\text{mean}}$ (section 2). c) Map of regression slopes (B) for the STF response regressed on the normalised SAM index response. B is consistently positive here, suggesting that models predicting the greatest increase in the SAM index also predict the lowest weakening of the AMOC. d) Map of the $R^2$ statistic based on the regressions in c. e) Map of regression slopes (B) for the STF response regressed on the initial $(1960-1999)_{\text{mean}}$ AMOC index. Models with the strongest initial AMOC also tend to show the greatest weakening of the AMOC. f) Map of the $R^2$ statistic based on the regressions in e. The areas bounded by white contours in c) and e) indicate where the results are inconsistent with the null hypothesis at the 95% confidence level.
Southern hemisphere maps of the multi-model mean response (2060-2099 minus 1960-1999) of 6 variables: a) Zonal (westerly) surface wind speed ($U_{surf}$). b) Barotropic streamfunction (STF BAROT). c) Northward Ekman transport per 2° latitude ($V_{ek}$). d) Precipitation minus evaporation (P-E). e) Salinity at 100m below the ocean surface (S). f) Sea-surface-temperature ($\Theta_{surf}$).

Maps of best fit regression line slopes (B) for regressions on the SAM index response in the southern hemisphere. The SAM index response has been normalised so that each map shows the pattern related to one standard deviation of the spread between the models. a) Zonal (westerly) surface wind speed ($U_{surf}$). b) Barotropic streamfunction (STF BAROT). c) Northward Ekman transport per 2° latitude ($V_{ek}$). d) Precipitation minus evaporation (P-E). e) Salinity at 100m below the ocean surface (S). f) Sea-surface-temperature ($\Theta_{surf}$). The areas bounded by white contours indicate where the results are inconsistent with the null hypothesis at the 95% confidence level.

Multi-model mean, zonally averaged Atlantic salinity for the period (1960-1999)$_{mean}$, showing the three major water masses in the Atlantic ocean.

Atlantic meridional cross sections of: a) Multi-model mean response (2060-2099 minus 1960-1999) in potential temperature ($\Theta$). b) Best fit regression line slopes (B) for the $\Theta$ response regressed on the SAM index response across model space. c) Multi-model salinity (S) response. d) Best fit regression line slopes (B) for the S response regressed on the SAM index response across model space. e) Multi-model potential density ($\rho$) response. f) Best fit regression line slopes (B) for the $\rho$ response regressed on the SAM index response across model space. The SAM index response has been normalised so that each map shows the pattern related to one standard deviation of the spread between the models. The areas bounded by white contours indicate where the results are inconsistent with the null hypothesis at the 95% confidence level.
Northward salt transport (TS) response at 30°S integrated over the depth and width of the Atlantic Ocean, plotted against a) the SAM index response and b) the change in the maximum AMOC streamfunction at 24°S (near to where the salt flux is calculated).

Northern hemisphere maps of the multi-model mean response (2060-2099 minus 1960-1999) of 6 variables: a) Zonal (westerly) surface wind speed ($U_{surf}$). b) Barotropic streamfunction (STFBAROT). c) Northward Ekman transport per 2° latitude ($V_{ek}$). d) Precipitation minus evaporation (P-E). e) Salinity at 100m below the ocean surface (S). f) Sea-surface-temperature ($\Theta_{surf}$).

Maps of best fit regression line slopes (B) for regressions on the SAM index response in the northern hemisphere. The SAM index change has been normalised so that each map shows the pattern related to one standard deviation of the spread between the models. a) Zonal (westerly) surface wind speed ($U_{surf}$). b) Barotropic streamfunction (STFBAROT). c) Northward Ekman transport per 2° latitude ($V_{ek}$). d) Precipitation minus evaporation (P-E). e) Salinity at 100m below the ocean surface (S). f) Sea-surface-temperature ($\Theta_{surf}$). The areas bounded by white contours indicate where the results are inconsistent with the null hypothesis at the 95% confidence level.
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Fig. 6. Atlantic meridional cross sections of: a) Multi-model mean response (2060-2099 minus 1960-1999) in potential temperature (Θ). b) Best fit regression line slopes (B) for the Θ response regressed on the SAM index response across model space. c) Multi-model salinity (S) response. d) Best fit regression line slopes (B) for the S response regressed on the SAM index response across model space. e) Multi-model potential density (ρ) response. f) Best fit regression line slopes (B) for the ρ response regressed on the SAM index response across model space. The SAM index response has been normalised so that each map shows the pattern related to one standard deviation of the spread between the models. The areas bounded by white contours indicate where the results are inconsistent with the null hypothesis at the 95% confidence level.
Fig. 7. Northward salt transport (TS) response at 30°S integrated over the depth and width of the Atlantic Ocean, plotted against a) the SAM index response and b) the change in the maximum AMOC streamfunction at 24°S (near to where the salt flux is calculated)
Fig. 8. Northern hemisphere maps of the multi-model mean response (2060-2099 minus 1960-1999) of 6 variables: a) Zonal (westerly) surface wind speed ($U_{surf}$). b) Barotropic streamfunction (STFBAROT). c) Northward Ekman transport per 2° latitude ($V_{ek}$). d) Precipitation minus evaporation (P-E). e) Salinity at 100m below the ocean surface (S). f) Sea-surface-temperature ($\Theta_{surf}$).
Fig. 9. Maps of best fit regression line slopes (B) for regressions on the SAM index response in the northern hemisphere. The SAM index change has been normalised so that each map shows the pattern related to one standard deviation of the spread between the models. a) Zonal (westerly) surface wind speed ($U_{\text{surf}}$). b) Barotropic streamfunction (STFBAROT). c) Northward Ekman transport per 2° latitude ($V_{\text{ek}}$). d) Precipitation minus evaporation (P-E). e) Salinity at 100m below the ocean surface (S). f) Sea-surface-temperature ($\Theta_{\text{surf}}$). The areas bounded by white contours indicate where the results are inconsistent with the null hypothesis at the 95% confidence level.