Exploring Mechanisms of Variability and Predictability of Atlantic Meridional Overturning Circulation in Two Coupled Climate Models

ROSS TULLOCH * AND JOHN MARSHALL *

Department of Earth, Atmospheric and Planetary Sciences,

Massachusetts Institute of Technology, 77 Massachusetts Avenue, Cambridge, MA

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Abstract

Multi-decadal variability in the Atlantic Meridional Overturning Circulation (AMOC) of the ocean is diagnosed in the NCAR Community Climate System Model (CCSM3) and the GFDL Coupled Model (CM2.1). Common diagnostic approaches are applied to draw out similarities and differences between the two models. An index of AMOC

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Corresponding authors' addresses:

Ross Tulloch, Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, 77 Massachusetts Avenue, Cambridge MA 02139. E-mail: tulloch@mit.edu

John Marshall, Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, 77 Massachusetts Avenue, Cambridge MA 02139. E-mail: jmarsh@mit.edu

variability is defined and the manner in which key variables covary with it is determined. In both models it is found that: (i) AMOC variability is associated with upper ocean (top 1 km) density anomalies (dominated by temperature) on the western margin of the basin in the region of the Mann Eddy with a period of about 20 years. These anomalies modulate the trajectory and strength of the North Atlantic Current. The importance of the western margin is a direct consequence of the thermal wind relation and is independent of the mechanisms that create those density anomalies.(ii) Density anomalies in this key region are part of a larger-scale pattern that propagates around the subpolar gyre and acts as a 'pacemaker' of AMOC variability. (iii) The observed variability is consistent with the primary driving mechanism being stochastic wind-curl forcing, with Labrador Sea convection playing a secondary role. Finally, 'toy models' of delayed oscillator form are fitted to power spectra of key variables and used to infer "quality factors" (Q-factors), which characterize the bandwidth relative to the center frequency and hence AMOC predictability horizons. The two models studied here have Q-factors of around 2, suggesting that prediction is possible out to about two cycles, which is likely larger than the real AMOC.

1. Introduction

The Atlantic Meridional Overturning Circulation of the ocean (AMOC) plays a central role in climate, transporting heat across the equator in to the North Atlantic basin. It is thought to modulate, for example, sea surface temperature (SST) and mediate sea ice extent in high northern latitudes (Kushnir 1994; Mahajan et al. 2011a). The AMOC in coupled climate models often exhibits pro-

nounced decadal variability that might be reflected in, and inherited by, other parts of climate system. Moreover, if decadal AMOC variability is associated with slow ocean processes which are predictable, then such predictability (if it exists) might be leveraged and exploited in coupled climate models used for climate change prediction and attribution — see Leetmaa and Marshall (2006); Hurrell et al. (2006) for a background discussion.

Currently there is intense interest in understanding the AMOC, documenting how it varies in observations (Kushnir 1994; Delworth and Mann 2000; Johns et al. 2011) and in models (Weaver et al. 1993; Delworth et al. 1993; Griffies and Tziperman 1995; Jayne and Marotzke 2001; Dai et al. 2005; Dong and Sutton 2005; Jungclaus et al. 2005; Danabasoglu 2008; Tziperman et al. 2008; Hawkins and Sutton 2009), and identifying the causes of that variability. Thus far it has not been possible to unambiguously identify a single driving mechanism and indeed there seems to be much model dependence in AMOC variability — see Marshall et al. (2001); Danabasoglu (2008); Yoshimori et al. (2010); Kwon and Frankignoul (2011); Sévellec and Fedorov (2011, and references therein).

AMOC variability diagnosed in models typically falls into one of the following types:

1. self-sustained internal (ocean-only) oscillations (Colin de Verdière and Huck 1999; Te Raa et al. 2004; Zhu and Jungclaus 2008; Buckley 2011).

2. a damped oceanic mode stochastically-excited by the atmosphere (Delworth et al. 1993; Griffies and Tziperman 1995; Delworth and Greatbatch 2000; Dai et al. 2005; Dong and Sutton 2005; Sévellec and Fedorov 2011; Frankcombe et al. 2009) and/or freshwater inter- actions with the Arctic (Jungclaus et al. 2005; Delworth et al. 1997; Olsen et al. 2008).

3. coupled atmosphere-ocean modes (Weaver and Valcke 1998; Eden and Willebrand 2001;

Oka and Hasumi 2006)

Some of the present uncertainty reflects the undoubted complexity of the processes at work. However, this is compounded by the fact that different diagnostic methods are employed by different investigators who come to the problem with differing viewpoints. Here we attempt to rationalize the situation somewhat. In Section 2 we diagnose AMOC variability in two different coupled climate models — the NCAR Community Climate System Model 3 (CCSM3) and the GFDL Coupled Model 2.1 (CM2.1). An index of AMOC variability is defined and the manner in which key variables covary with it is used to draw out aspects of underlying mechanisms and similarities and differences between the two models. We draw attention to the role of subsurface density anomalies that abut the continental shelf on the western margin of the Atlantic basin in the region of the Mann Eddy. This region is located at the latitude at which the mean AMOC is a maximum and where it exhibits maximum variability in the two models. In Section 3 we fit 'toy models' of delayed oscillator form to observed spectra of key variables and use them to infer 'Q-factors', characterizing the bandwidth relative to the center frequency, and hence AMOC predictability horizons. Finally, in Section 4, we conclude, drawing out aspects that might pertain to the real, rather than the modeled, ocean.

2. Variability in the Atlantic Ocean in two coupled climate models

Here we explore the variability of Atlantic Meridional Overturning Circulation (hereafter AMOC) in 500-year control integrations of the NCAR Community Climate System Model 3 (CCSM3) and

the GFDL Coupled Model 2.1 (CM2.1). The detailed configurations of each model are described in Appendix A. AMOC variability has been studied before in these two models, as we now briefly describe.

In Danabasoglu (2008), deep convection in the Labrador Sea region was identified as a key precursor to decadal AMOC variability in CCSM3, and complicated interactions between the MOC and deep water formation, the North Atlantic Oscillation and the subpolar gyre circulation are identified as drivers of MOC variability. Yoshimori et al. (2010) analyzed a low resolution version of CCSM3 and argued that variability involves an interaction between advection of salt and surface density in the subpolar gyre, with a connection to the North Atlantic Oscillation (NAO). A positive feedback involving eddy activity in the atmosphere was invoked. Meanwhile, Kwon and Frankignoul (2011) identified regimes of high and low AMOC variability in CCSM3. The first regime (studied by Danabasoglu 2008) has a very regular 20 year oscillation and the second regime has red noise-like variability. They concluded that the variability in the second regime could be understood as a damped oceanic mode that is stochastically-excited by the NAO. They also noted that a delayed positive feedback caused by changes in horizontal ocean circulation increased AMOC persistence from 5 to 10 years.

Zhang (2008) noted that AMOC variability in CM2.1 is associated with a dipole in subsurface temperature anomalies. A positive AMOC is associated with negative temperature anomalies in the Gulf Stream region and slightly weaker positive temperature anomalies in the subpolar gyre, indicating that a strengthening of the AMOC in CM2.1 is associated with a weakening of the subpolar gyre. Zhang et al. (2011) argued that North Brazil Current transport lags Labrador Sea convection in CM2.1 and is a useful indicator for AMOC variability. These two studies imply that freshwater fluxes likely play a role in generating AMOC variability. On the other hand, Frankcombe et al.

(2009) argued that the dominant timescale of North Atlantic variability in CM2.1 is determined by the time for subsurface temperature anomalies to propagate westwards as baroclinic Rossby waves.

Results from previous studies on CCSM3 and CM2.1 indicate that a number of factors may affect AMOC variability, and that there is no single underlying mechanism across models. The prevailing view appears to be that AMOC variability is excited by external NAO forcing in CCSM3 and freshwater fluxes in CM2.1. Thus far there has been little attempt to draw parallels between the two models and identify common driving mechanisms in them. This is the goal of the present study. We argue that, in both models, density anomalies are created in the subpolar gyre and affect AMOC via the thermal wind relation upon reaching the western boundary.

We begin by defining an index of MOC variability and go on to study how other aspects of the circulation, such as temperature (T), salinity (S) and potential density (ρ) anomalies, covary with it. Much use is made of composite property maps defined at various leads and lags referenced to a (HI)-(LO) MOC index, which is defined in Section 2a. Emphasis is placed on density anomalies on the western margin of the gyre in the region of the Grand Banks, which play a key role in pacing MOC variability in both models. We conclude the section by drawing together aspects of the variability that are common between the two models. This is used to motivate the construction of simplified 'toy models' described in Section 3.

115   **2a.** *MOC Variability*

116   Fig. 1(a,d) shows contours of the time averaged MOC (in Sv) in CCSM3 and CM2.1.[1] In both

117   models the maximum overturning occurs just north of 40°N at a depth of roughly 1 km. Time

118   series of annual-mean MOC averaged over the regions indicated by the black box (roughly from

119   35° to 50° and 1800 m to 500 m depth) in each model are shown in Fig. 1(b,e). Years that are more

120   than one standard deviation above (below) the mean are marked with x's (squares). These years

121   will now be used to create composite variables based on a (HI)-(LO) MOC index, *i.e.,* composites

122   are the average of a given variable over years when the MOC is (HI) minus the average when the

123   MOC is (LO). This technique was used, for example, in Czaja and Marshall (2001) to study the

124   evolution of SST anomalies in the Atlantic based on an SST dipole index.

125   The (HI)-(LO) MOC pattern is shown in Fig. 1(c,f). It has a maximum magnitude of 6.8 Sv

126   in CCSM3 and 4.6 Sv in CM2.1 and is located near the black box at zero lag. The patterns are

127   slightly deeper than the mean MOC, which is consistent with the first empirical orthogonal function

128   (EOF1) (see Danabasoglu 2008).

129   **2b.** *Covarying temperature signals*

130   Figs. 2(a,d) show power spectra of the MOC indices (red lines) in Fig. 1 for both CCSM3 and

131   CM2.1. Also plotted are power spectra of a temperature index (T), defined as the upper 1 km tem-

132   perature averaged over the gray-boxed region near the Grand Banks in Figs. 2(b,c,e,f). The gray

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[1]Only the explicitly calculated, Eulerian component of the MOC is considered here, as in Danabasoglu (2008). In

CM2.1, the parameterized bolus contribution has a negligible effect on the MOC spectrum since its variance is less

than 1% of the Eulerian contribution.

133 shading indicates 95% confidence limits of the MOC for a χ^2 distribution. The significance of this

134 boxed region will become more clear in Sections 2d and 2e below. In both models, peaks in vari-

135 ability are evident at approximately 20 years. Notice that the spectral content of the MOC-index

136 and the $T-$index are very similar, indicating that the MOC and T indices are varying together,

137 with perhaps a delay between them. The green lines show ω^{-2} power laws: on subdecadal time-

138 scales the T signal is likely to be integrating a white noise process, consistent with Frankignoul

139 and Müller (1979). Later, in Section 3, we will model the decadal peak as a delayed oscillator

140 using ideas in Marshall et al. (2001).

141 Figs. 2(b,e) show composite maps of T based on the (HI)-(LO) MOC index defined in Fig. 1.

142 Regions where the (HI)-(LO) composites are not statistically significant at the 95% confidence

143 level with the two-tailed Student's t-test are shaded white. Note that the temporal degrees of

144 freedom are estimated using Effective Sample Size diagnostic of Bretherton et al. (1999). The

145 spatial patterns of T at years with high MOC minus T in years with low MOC closely matches the

146 first Empirical Orthogonal Function (EOF1) of T shown in Figs. 2(c,f). Thus the primary mode of

147 variability in T is nearly in phase with MOC. Cross-correlation analysis, not shown, reveals that

148 (minus) the boxed $T-$index lags MOC by 2 years in CCSM3, with cross-correlation of about 0.47

149 (compared to a 95% confidence level of about 0.2), while in CM2.1 the $T-$index is in phase with

150 MOC with correlation 0.7 (compared to a 95% confidence level of about 0.185). As described in

151 Section 2f, the $T-$index delay in CCSM3 is likely due to deep convection in the Labrador Sea,

152 which provides a slow pathway for subsurface density anomalies near 1 km depth to reach the

153 Grand Banks region.

154 In both models, most of the variability of the MOC-index is associated with temperature dom-

155 inated density anomalies on the western boundary near the Grand Banks, *i.e.,* the blue regions

156 within the gray boxes in Fig. 2, which will be described in more detail in Section 2d. The black

157 contours in Figs. 2(b,c,e,f) indicate the time-mean zero wind stress curl lines, roughly marking the

158 gyre boundaries. The close correspondence between the maximum temperature variations and the

159 zero-wind stress curl line is indicative of a possible role of wind forcing, because the greatest part

160 of the wind stress curl variability lies along the zero wind stress curl line.

2c. *Covarying density signals*

162 Fig. 3 shows composite maps of potential density averaged over the top 1 km using the (HI)-(LO)

163 MOC index in CCSM3. Fig. 3(b) is computed by subtracting the high-index composite from the

164 low-index composite. The (HI)-(LO) map in Fig. 3(b) is very similar to the map in Fig. 2(b) (but

165 with opposite sign), indicating that density anomalies at high and low MOC states are dominated

166 by temperature anomalies. The corresponding (HI)-(LO) salt map (not shown) has very similar

167 patterns and signs as Fig. 2(b), showing that T and S are compensated, except in the Labrador Sea

168 region (as noted by Danabasoglu 2008). The most pronounced feature of the density pattern (blue

169 region in Fig. 3b) forms near the zero wind stress curl line and the mid-Atlantic ridge. The '5 years

170 earlier' map in Fig. 3(a) is obtained by taking the average of $(\text{HI} - 5)$ years minus the average

171 of $(\text{LO} - 5)$ years. It shows advection of a positive density anomaly around the subpolar gyre.

172 The '5 years later' map in Fig. 3(d) shows the opposite phase, with a negative density anomaly

173 beginning to advect cyclonically around the subpolar gyre. Also, the positive density anomaly that

174 previously propagated around the subpolar gyre has merged with another positive density anomaly

175 that formed in-situ off of the Grand Banks (*i.e.,* inside the boxed region in Fig. 3b). This merging

176 of positive density anomalies is also visible in the meridional section shown in Fig. 5(c). Labrador

Sea anomalies subduct and propagate southward before merging with a shallower density anomaly in the Grand Banks region.

Fig. 4 shows maps of potential density based on the composite (HI)-(LO) MOC index in CM2.1, analogous to Fig. 3 for CCSM3. In CM2.1, as in CCSM3, the composite density map at 0 years (Fig. 4b) closely matches the composite temperature map in Fig. 2(d): Temperature dominates, with salinity compensating (not shown) everywhere except in the region of deep convection in the Labrador Sea, where salt anomalies dominate at 0 lag. Again, the largest density anomalies are found along the zero wind stress curl line, but they originate further north near the mid-Atlantic ridge and propagate both west into the Labrador Sea and southwest into the region off the Grand Banks.

2d. *Role of density anomalies on the western boundary*

Buckley (2011) showed that MOC variability occurs almost entirely on the western boundary in coupled aqua-planet simulations with meridional continental barriers. In one of their model configurations, with flat bottom bathymetry, anomalies generated near the eastern boundary propagate westward and trigger MOC variations when they reach the western boundary, in accord with the thermal wind relation. The temperature anomalies shown in Fig. 2 form in different locations in the two more realistic models studied here, but all models suggest that density anomalies propagate around the subpolar gyre and influence the MOC when they reach the western boundary.

Here, motivated by that study, we investigate the role of density anomalies on the western boundary and their association with MOC anomalies. Fig. 5 shows composites of zonal and meridional cross-sections of meridional velocity and density in CCSM3 (top row) and CM2.1 (bottom

198 row), from the regions indicated by the gray dashed lines in panel d of Figs. 3 and 4. The zonal

199 cross-sections are averaged from $40°$N to $45°$N, which is approximately the latitude range where

200 the MOC is largest. The meridional cross-sections are averaged from $50°$W to $40°$W, encompass-

201 ing the longitude where the deep western boundary current flows offshore of the Grand Banks

202 and where density anomalies associated with MOC variability are found. Fig. 5(a) shows that the

203 largest transport differences between high and low MOC years occurs around $40°$W, where both

204 the Gulf Stream and the deep western boundary current are anomalously strong. The tripole pattern

205 of meridional velocity in Fig. 5(a) is consistent with the density dipole present in the zonal section

206 of Fig. 5(b). However, as is evident from a consideration of the thermal wind equation (see below)

207 the positive density anomaly on the western boundary is associated with the major part of MOC

208 variability.

209 The bottom row of Fig. 5 shows zonal and meridional composite cross-sections of meridional

210 velocity and density in CM2.1, averaged over the same latitude and longitude bands. Analogously

211 to CCSM3, the transport anomalies in Fig. 5(d) are centered over $40°$W and are a maximum when

212 the Gulf Stream and deep western boundary current are strong. Again, there is also a clear positive

213 density anomaly on the western boundary in Fig. 5(e).

214 The anomalous meridional transport can be related to zonal density gradients through the ther-

215 mal wind relation (Hirschi and Marotzke 2007), thus:

$$f\frac{\partial v}{\partial z} = -\frac{g}{\rho_0}\frac{\partial \rho}{\partial x} \qquad \Rightarrow \qquad \overline{v}^x(z') = \int_{x_W}^{x_E} v\,dx = \frac{g}{\rho_0 f}\int_{-H}^{z'}\left[\rho(WB) - \rho(EB)\right]dz. \quad (1)$$

216 where ρ_0 is the mean density, f is the Coriolis parameter, g is the acceleration due to gravity

217 and $\rho(WB)$ and $\rho(EB)$ are the density anomalies on the western and eastern boundaries. On

218 integrating vertically we obtain an anomalous overturning streamfunction of

$$\psi_\rho(z) \approx -\int_{-H}^{z} \left[\overline{v}^x(z') - \frac{1}{H}\int_{-H}^{0} \overline{v}^x dz \right] dz', \tag{2}$$

219 where the vertical average of $\overline{v}^x$ has been subtracted out to ensure that $\psi_\rho(z = 0) = \psi_\rho(z =$

220 $-H) = 0$.

221   In Fig. 6 we compare the MOC variability $\text{MOC}(HI) - \text{MOC}(LO)$ (solid lines) with the

222 anomalous transport computed from the density anomalies on the western boundary $\psi_\rho(WB)$

223 (dashed lines) in the two models. The contribution of density anomalies on the eastern bound-

224 ary $\psi_\rho(EB)$ is negligible in CCSM3 and is much smaller than $\psi_\rho(WB)$ in CM2.1 (as indicated in

225 Fig. 6b). The MOC variability $\text{MOC}(HI) - \text{MOC}(LO)$ is defined as the average MOC in 'high'

226 years minus the average MOC in 'low' years averaged from $40°$N to $45°$N, while $\psi_\rho(WB)$ is com-

227 puted using Eq. (2) applied to the density fields shown in Fig. 5(b) and 5(e). The dashed lines in

228 Fig. 5(b) and 5(e) indicate regions over which density was zonally averaged to estimate $\psi_\rho(WB)$,

229 that is from the shelf out to the dashed line.

230   The close match between $\text{MOC}(HI) - \text{MOC}(LO)$ and $\psi_\rho(WB)$ in both models is perhaps not

231 unexpected in view of the thermal wind equation. As noted above, this is also a property of the

232 coupled, idealized models studied in Buckley (2011). It highlights the close association between

233 density anomalies on the western boundary and MOC variability. Understanding how those density

234 anomalies are created, how they arrive at the western boundary and what sets their timescale of

235 variability, is key to understanding how the MOC varies.

2e. *Role of the Mann Eddy*

As shown in Figs. 2-4, MOC variability is associated with upper ocean, temperature-dominated density anomalies at the western boundary near the Grand Banks. Fig. 7 shows mean currents and (HI)-(LO) composites at 300 m depth in the two models. The (HI)-(LO) composites of each model reveal that in association with a strong MOC there is anomalously cyclonic circulation in the boxed region, and a somewhat enhanced Gulf Stream and Labrador Current. These cyclonic anomalies are in geostrophic balance with the positive buoyancy anomalies shown in Figs. 3(b) and 4(b). We believe it is significant that these cyclonic anomalies are at the location of the respective model's representation of the Mann Eddy. In the ocean, the Mann Eddy is a persistent anticyclone swirling at the end of the separated Gulf Stream to the southeast of the NAC, located approximately between $38°N–45°N$ and $40°W–50°W$ (Mann 1967; Meinen et al. 2000). This is the region where the deep expression of the southward flowing Labrador Current abuts the northward flowing Gulf Stream. The variability of the Mann Eddy likely modulates the trajectory of the NAC and the strength of the MOC. We hypothesize that the cyclonic circluation within the boxed regions of the (HI)-(LO) composites in Fig. 7(b,d) imply that a weakened Mann Eddy is associated with an enhanced MOC.

2f. *Role of Labrador Sea Convection*

In CCSM3, there is a clear connection between density anomalies that propagate around the sub-polar gyre, precondition convection, and result in density anomalies that later appear in the $40°N$ to $45°N$ latitude band (see Fig. 5b). Fig. 8(a) shows (HI)-(LO) composite maps of March mixed layer depth (MLD). The first EOF of MLD variability shown in Fig. 8(b) is very similar to the (HI)-(LO) composite maps. Cross-correlations of T, S, ρ (at 150m depth in the box in the region

257 of maximum MLD variability) against MLD (in the same box) and the MOC index defined in

258 Fig. 1 are shown in Fig. 8(c,d). Density anomalies are in phase with the MLD anomalies and so it

259 seems clear that the propagation of density around the subpolar gyre is preconditioning Labrador

260 Sea deep convection — see the review in (Marshall and Schott 1999). Fig. 8(d) shows that density

261 anomalies in the Labrador Sea region lead MOC anomalies (see also Danabasoglu 2008). The

262 propagation of deep density anomalies from the Labrador Sea to the region of maximum MOC

263 can be seen coming from north of $50°$ in Fig. 5(c). However, a significant fraction of the density

264 anomaly at $42°$N does not originate in the Labrador Sea.

265 In CM2.1, the connection between MOC variability and deep convection in the Labrador Sea

266 is less clear. The principle component time series of EOF1 of MLD does covary with MOC in-

267 dex however its spatial pattern is a large dipole covering the Labrador Sea region (not shown).

268 Consequently, cross-correlations between (T, S, ρ) in the Labrador Sea (northernmost boxed re-

269 gion shown in Fig. 8) and the maximum MOC (*i.e.,* the index defined in Fig. 1) are insignificant.

270 Similarly, cross-correlations between (T, S, and ρ) and MLD are not significant since the north-

271 ern box overlaps the EOF-1 dipole. Fig. 5(f) also indicates that MOC variability is mainly due to

272 near-surface density anomalies south of $50°$N, with little indication of a role for convection north

273 of $50°$N.

274 In summary, Labrador Sea convection does seem to play some role in creating density anoma-

275 lies off George's Bank in CCSM3, but there is little evidence that it is creating density anomalies

276 there in CM2.1. In both models density anomalies off George's Bank set the strength of the MOC

277 through thermal wind, and are controlled by the advection of density anomalies around the subpo-

278 lar gyre.

3. Idealized 'toy model' of MOC variability

3a. *Construction of an 'oscillator model'*

In the previous section we showed that MOC variability covaries with temperature-dominated density anomalies on the western boundary of the subpolar gyre. The dominant timescale of the variability appears to depend on the time it takes for the density anomalies to propagate around the subpolar gyre and reach the western boundary. We hypothesize, as do Kwon and Frankignoul (2011) and Danabasoglu (2008), that such anomalies could be energized by stochastic atmospheric forcing. In this section we apply the ideas of Marshall et al. (2001) to formulate a delayed oscillator 'toy model' of the MOC variability. This is a modification of the canonical model of the interaction of ocean with the atmosphere due to Frankignoul and Hasselmann (1977, in the following FH77):

$$\frac{\partial T}{\partial t} = \mathcal{F} - \lambda T, \tag{3}$$

where T denotes a near-surface temperature (FH77 assumed it to be SST), λ is a damping timescale (of order of a season) and $\mathcal{F}$ is the surface forcing (Ekman pumping and turbulent heat flux) associated with atmospheric variability. In FH77 $\mathcal{F}$ is taken to be a white noise process, so that Eq. (3) predicts a red spectrum for T on timescales shorter than λ^{-1}, and a flattening on longer timescales. This is obviously not a good model of the processes that lead to the spectra shown in Fig. 2. which exhibit a pronounced spectral peak at (see below) roughly the time it takes a baroclinic Rossby wave to propagate across the basin.

We therefore make use of a simple elaboration of Eq. (3) which is capable of capturing a spectral peak through a delayed oceanic response of T to external meteorological forcing. Accordingly we modify Eq. (3) with a delay term, D, thus, to represent ocean circulation:

$$\frac{\partial T}{\partial t} = -\tau - \lambda T - DT\left(t - t_d\right), \tag{4}$$

where

$$\tau = \alpha\mathcal{N} - fT. \tag{5}$$

In our interpretation of Eq. (4), T denotes the strength of upper ocean thermal anomalies on the western boundary (for example in the box in Fig. 7), τ is the amplitude of the wind stress variability (in temperature units) blowing over the basin which is assumed to be decomposed into a stochastic component $\mathcal{N}$ and, possibly, an SST—induced feedback component (fT). Key model parameters are (they are discussed in much more detail in Marshall et al. 2001):

- α: scaling of stochastic wind stress $\mathcal{N}$ into a temperature tendency,

- λ: damping due to air-sea interaction,

- f: feedback of T on the wind stress pattern,

- D and t_d measure the strength and time-delay of the feedback of ocean circulation on T.

We present the delay term in Eq. (4) as a representation of advective and wave ocean processes that, on long timescales, play a role in setting T. Clearly it is a highly symbolic characterization of the processes at work in the ocean and ocean models. However, it should be noted that, as shown in Marshall et al. (2001), a term of this form arises from a consideration of time-dependent Sverdrup dynamics and thermohaline circulation dynamics. Here we are agnostic as to the detailed processes that control D and t_d and, instead, adjust these model parameters to best fit model spectra.

For $f = 0$, the solution to Eq. (4) in the Fourier domain is given by

$$\hat{T} = -\frac{\alpha\hat{\mathcal{N}}}{i\omega + De^{-i\omega t_d} + \lambda}, \tag{6}$$

where $|\hat{\mathcal{N}}|^2 = 1$, and $T(t) = \int \hat{T}(\omega)e^{i\omega t}d\omega$. Fig. 9 shows the (a) spectra and (b) "free solutions"

for $D = 0, 0.1, 0.2$ and 0.3 years^{-1} with $\alpha = 1$ K·m^2·N^{-1}, $t_d = 8$ years, and $\lambda = 2\pi/32$ years^{-1},

as in Plate 1 of Czaja and Marshall (2000). The free (or unforced) solutions, are obtained by setting

$\alpha = 0$, initializing with a cosine function for $t < 0$, and then allowing them to decay away in time,

thus:

$$T_{\text{free}} \propto e^{-\gamma t/2} \cos(\omega_0 t), \tag{7}$$

where ω_0 measures the frequency of the oscillation and γ the decay rate. The parameter γ in Eq. (7)

is obtained by a least squares fit over the first few oscillations of the free solutions. Decay rates

in Fig. 9 are $\gamma = 2\pi/16, 2\pi/13.5, 2\pi/29.5$, and $2\pi/132.5$ (years)$^{-1}$ for $D = 0, 0.1, 0.2$ and 0.3

(years)$^{-1}$ respectively. From these decay rates and ω_0 we can define a Q-factor:

$$q = \frac{\omega_0}{\gamma}. \tag{8}$$

The Q-factor measures the number of distinct oscillations and is approximately $0, 1, 2.5$ and 11.5

in the four cases. The Q-factor can also be measured directly from the spectral density of T as the

peak frequency divided by the bandwidth of the range of frequencies for which the energy is half

its peak value.

3b. *Fitting the oscillator model to the 'observed' spectra*

To fit the delayed oscillator model to observed spectra it is useful to consider two extreme fre-

quency limits implied by Eq. (6) as follows:

$$\lim_{\omega \to 0} |\hat{T}|^2 = \left(\frac{\alpha}{D + \lambda}\right)^2 \tag{9}$$

332 and

$$\lim_{\omega \to \infty} |\hat{T}|^2 = \frac{\alpha^2}{\omega^2}. \tag{10}$$

333 We first solve for α in the $\omega \to \infty$ limit, then solve for $D + \lambda$ using the $\omega \to 0$ limit. Note that the

334 time delay t_d is tightly constrained by the frequency of the spectral peak because resonance occurs

335 when $\omega t_d = \pi/2$, giving an oscillation period of $4t_d$ in the absence of damping (*i.e.,* $\lambda = 0$). Also

336 note that D determines the height of the spectral peak, and λ acts as a limiter on growth towards

337 low frequencies.

338   Figs. 10(a,b) show normalized spectra (blue curves) for CCSM3 and CM2.1 and the recon-

339 structed spectra using a purely statistical AR_{15} least squares fit (black curves) and the delayed

340 oscillator model in Eq. (4) (red curves) fitted manually using the statistical fit as a guide[2]. The

341 fitted delayed oscillator parameters in CCSM3 are

342   • $\alpha = 0.6$, $t_d = 7$ years, $D = 0.17$ years^{-1} and $\lambda = 0.157$ years^{-1},

343   while in CM2.1 they are

344   • $\alpha = 0.4$, $t_d = 5$ years, $D = 0.2$ years^{-1} and $\lambda = 0.063$ years^{-1}.

345   From these parameters we can deduce ω_0 and γ in Eq. (7) and then compute the Q-factor in

346 CCSM3:

347   • $\omega_0 = 2\pi/24$ years^{-1}, $\gamma = 2\pi/43$ years^{-1} $\Rightarrow q = 1.8$

348   and for CM2.1:

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[2]Our delayed oscillator model has only three free parameters, yet it is compared to a higher-order autoregressive model here

349   • $\omega_0 = 2\pi/21.5 \text{ years}^{-1}$, $\gamma = 2\pi/38 \text{ years}^{-1} \Rightarrow q = 1.76$.

350   We note in passing that these fitted delay times t_d are broadly consistent with the transit time

351 of Rossby waves in the North Atlantic at 40°N. Using data from Tulloch et al. (2009), we estimate

352 the observed Rossby wave phase speed in the eastern Atlantic to be about 2 cm·s^{-1} from 40°N to

353 50°N, which gives transit times of the order of 5 to 10 years. Also, as pointed out by Kwon and

354 Frankignoul (2011), the regime from years 51-350 in CCSM are more predictable than all of the

355 years. We estimate the Q-factor for this range of years to be about $q = 3$.

356 **3c.** *Implications for predictability*

357 The delayed oscillator models fit the spectral peak, some of the higher harmonics, and the low

358 frequencies quite well, indicating that they can perhaps be used to make predictions. Here we

359 show a sample prediction from such a fitted toy model and compare it with a purely statistical

360 prediction, assuming perfect initial conditions. We simply train our toy model on the MOC time

361 series, but bear in mind that a real prediction would require training our model against a subsurface

362 temperature fingerprint, as done in Mahajan et al. (2011b).

363   Figs. 10(c,d) show ensemble mean predictions in CCSM3 and CM2.1 with the model param-

364 eters given above. To create the predictions, ensembles for the statistical and delayed oscillator

365 models were initialized with the MOC from the 15 years preceding the starting year, which is

366 denoted in the figure by vertical dashed lines. As time progresses, the ensemble means decay

367 away because the ensemble members become uncorrelated. The predictions shown here are the

368 ensemble means scaled by those decay rates ensuring that the magnitudes of the predictions re-

369 main roughly constant in time. Both the statistical (AR_{15}) and delayed oscillator predictions track

370 the MOC for a couple of oscillations before losing the phase, consistent with Q-factors of about 2.

371 As expected from their power spectra, the statistical and delayed oscillator predictions are almost

372 identical for the first couple of cycles, and then diverge as either fast timescale errors accumulate

373 or long timescale errors start to appear. However the quality of the predictions depend on how rep-

374 resentative the predicted years are to the average behavior over the 500 year time series. Note that

375 the predictions shown in Fig. 10 correspond to instances where the prediction tracks MOC longer

376 than suggested by the model's Q-factors. Other initial conditions result in predictions which lose

377 the phase of the MOC within a few years.

378 Our results are broadly in accord with those of Mahajan et al. (2011b), who attempted to pre-

379 dict AMOC variations using an autoregressive model (AR2) tuned to the subsurface temperature

380 anomalies, or "AMOC fingerprints", defined in Zhang (2008). They predicted a drop in AMOC

381 strength in the few years following its peak in 2005. This was based on the fidelity of their AR2

382 model in hindcasts of observed subsurface temperature anomalies and the close connection be-

383 tween AMOC variability and subsurface temperature variability in CM2.1. Mahajan et al. (2011b)

384 also found that hindcasts from their AR2 model show comparable skill whether they are trained

385 on 500 years or only 50 years of CM2.1 data, although they note that the 20 year spectral peak

386 in CM2.1 is not robust across models or the observations. Msadek et al. (2010) also argued that

387 AMOC in CM2.1 is predictable for up to 20 years.

388 **4. Summary and Discussion**

389 By using the same diagnostic tools to study variability in two coupled climate models, we have

390 identified aspects that are common to them both. In particular we find that:

(i) AMOC variability is associated with upper ocean (top 500 m to 1 km) density anomalies on the western margin of the basin in the region of the Mann Eddy. These anomalies modulate the trajectory and strength of the North Atlantic Current. The importance of the western margin is a direct consequence of the thermal wind relation and is independent of the mechanism that creates those density anomalies.

(ii) Density anomalies in this key region are part of a larger-scale density pattern that propagates around the subpolar gyre. These act as a 'pacemaker' of AMOC variability.

(iii) Density anomalies are dominated by temperature, with salinity playing a lesser role.

(iv) The timescale of AMOC variability is broadly consistent with the inherent timescale set by the transit time of first baroclinic mode Rossby waves at 45°N.

(v) Although we are unable to identify how the density anomalies are created, they appear to emanate along the zero-wind-stress curl line separating the subtropical and subpolar gyres. It is here that modeled (NAO) wind variability is a maximum. We suspect that this external wind forcing gives rise to thermocline undulations as in Frankignoul et al. (1997) which, when they strike the western boundary, spin up the AMOC. A hierarchy of ocean-only and coupled experiments would be required to confirm or reject this conjecture.

(vi) Density anomalies entering regions of deep convection (e.g. the Labrador Sea) can act to preconditions convection. This is clearly evident in CCSM3 but less important in CM2.1. Deep, convectively-formed anomalies can then propagate southward on the western boundary current to the region of the Mann Eddy and thence induce changes in the AMOC.

(v) Fitting a delayed oscillator model to the spectra found in the two models suggests that AMOC variability has a Q-factor of around 2, implying predictability out to a couple of cycles or so. This can be used as a minimalist model against which more complex predictability systems

414 can be compared.

415 Finally, it should be noted that an association between MOC variability and density anomalies

416 on the western margin of the gyre (off-shore of the Grand Banks in the region of the Mann Eddy)

417 is, in retrospect, not surprising. Zonal and vertical integration of the thermal wind equation links

418 changes in geostrophic contributions to MOC to those of depth integrated buoyancy differences

419 between the zonal endpoints. To the extent that anomalies on the eastern boundary are much

420 smaller than anomalies on the west, anomalies on the west become the dominant contributor to

421 MOC variability on decadal timescales. This could explain why, for example, MOC variability

422 is so sensitive to model formulation, both between models and in the same model when changes

423 are made to its resolution, overflow parameterizations etc. For example Danabasoglu et al. (2011)

424 and Yeager and Danabasoglu (2011) describe AMOC variability in CCSM4 and find significant

425 sensitivity to how Nordic Sea overflows are parameterized. The dynamics of the ocean in the

426 region of the Mann eddy is a 'cross roads' where potential vorticity anomalies created in distant

427 parts of the basin are brought together and interact. This is indeed a very complex region and will

428 be sensitive to model resolution and parameterization.

429 5. Acknowledgements

APPENDIX A

Climate models

The NCAR CCSM3 coupled model is described in detail in Danabasoglu (2008). It uses the Community Atmosphere Model 3 (CAM3), the Community Land Model Version 3 (CLM3), the Parallel Ocean Program version 1.4 (POP1.4) and the Community Sea Ice Model (CSIM). The atmospheric model is at T85 resolution with 26 vertical levels, and the ocean model has nominal $1°$ horizontal resolution and 40 vertical levels. Here we analyze years 101-600 of the 700 year present day (1990) control integration experiment b30.009.

The GFDL 2.1 coupled model, described in Delworth et al. (2006), has $2.5° \times 2°$ horizontal resolution and 24 vertical levels in the atmosphere, and uses the Modular Ocean Model 4 (MOM4), which has nominal $1°$ horizontal resolution and 50 vertical levels. Here we analyze years 1-500 of the pre-industrial (1860) control experiment 1860-D4.

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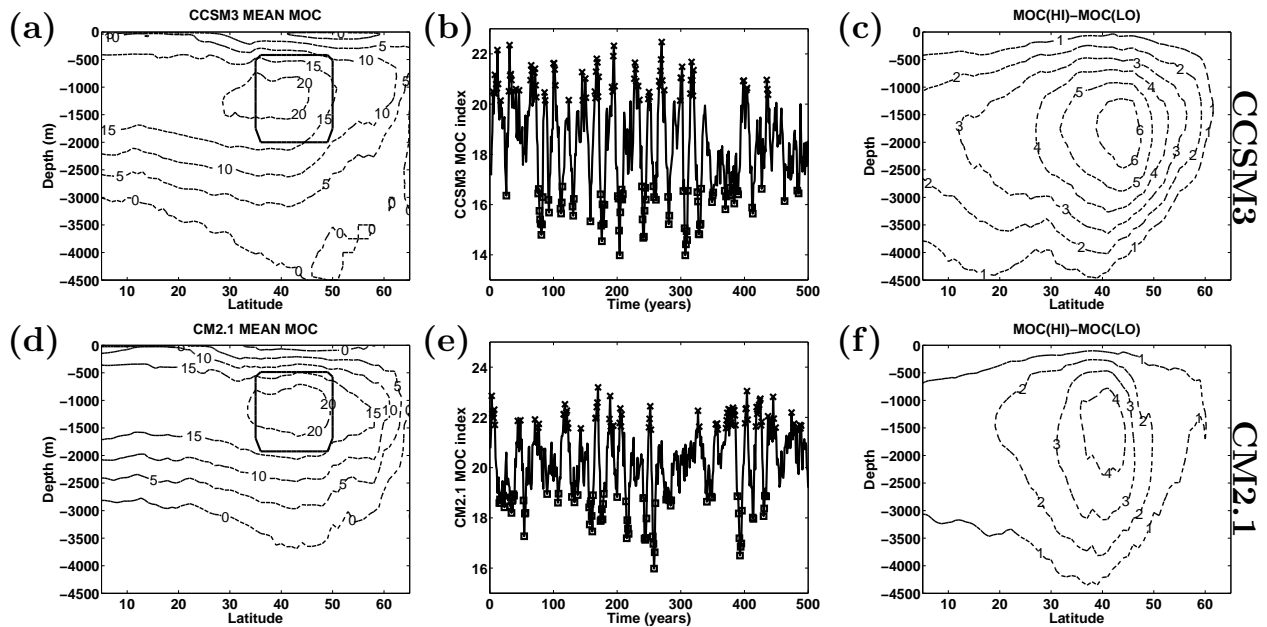569   as an ocean internal mode. *Clim. Dyn.*, doi:10.1007/s00382-008-0383-9.

Figure 1: (a) Eulerian-mean MOC (in units of Sv) in the Atlantic in CCSM3 and (b) time series of the MOC index, defined as the average MOC from $35°$N to $50°$N and 1800 m to 500 m depth, as indicated by the black box in (a). Years marked by x's (squares) denote years that are more than one standard deviation above (below) the time mean. These are used to construct composite maps. (c) The MOC averaged over years marked by x's minus the MOC averaged over years marked by squares, *i.e.,* the MOC(HI)-MOC(LO). The same variables are plotted for CM2.1 in (d), (e) and (f).

Figure 2: (a) Normalized power spectra of the MOC index (red), temperature (K) averaged over the top 1 km in the gray-boxed regions shown (b) and (c) (black), and the first principal component of temperature (K) in the top 1 km (blue) in CCSM3. The green line indicates a -2 spectral slope and the gray shading is a 95% confidence interval. (b) Composite (HI)-(LO) map of temperature averaged over the top 1 km, obtained by subtracting the low-index composite from the high-index composite, years marked by x's and squares respectively in Fig. 1(b). White shading indicates regions that are not significant at the 95% confidence level. (c) First EOF of temperature averaged over the top 1 km. The black contour indicates the time-mean zero-wind-stress-curl line. The same variables are plotted for CM2.1 in (d), (e), (f), see text for details.

Figure 3: Composite maps of potential density (kg·m^{-3}) averaged over the top 1 km using the index described in Fig. 2 for CCSM3. (a) The 5-years-earlier map is obtained by subtracting the (low index − 5-years) composite from the (high index − 5-years) composite. (b) As in (a) but at the composite index years, (d) as in (a) but 5 years after the composite years, (c) as in (a) but 10 years after the composite years. Thus, the anomalies evolve in time from the top left to the bottom left. White shading indicates regions that are not significant at the 95% confidence level.

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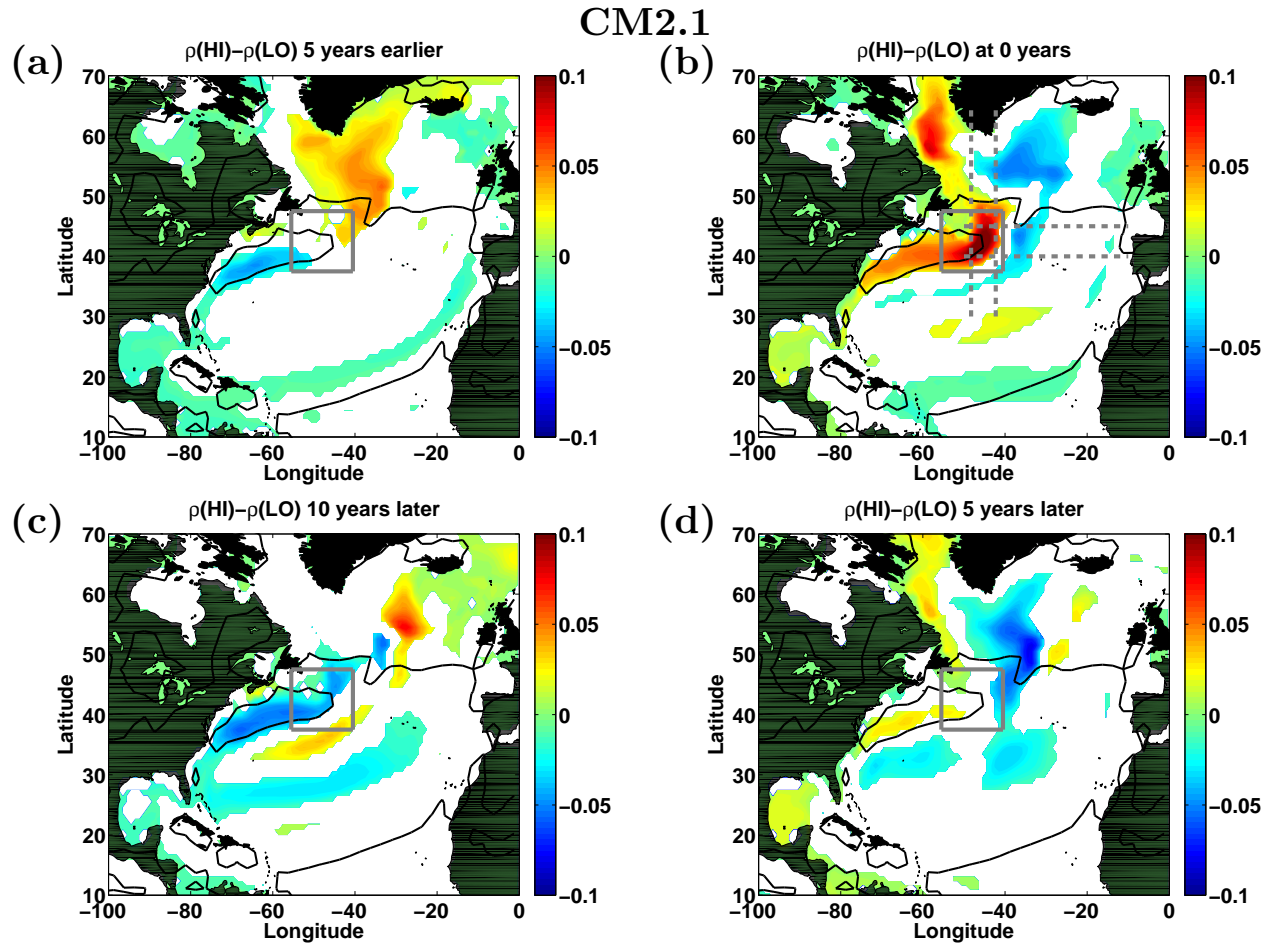
Figure 4: Same as Fig. 3 but for CM2.1.
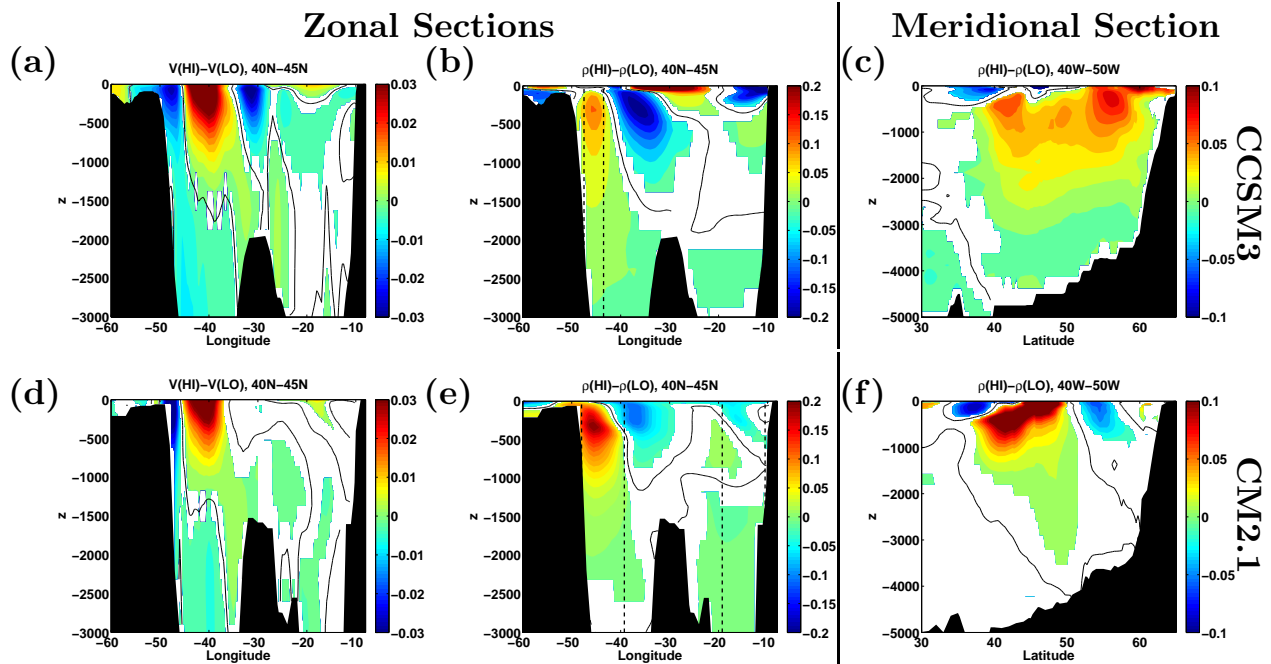
Figure 5: Composite zonal and meridional cross-sections of meridional velocity (m·s^{-1}) and density (kg·m^{-3}) in CCSM3 (top row) and CM2.1 (bottom row). The zonal cross-section averaged from 40°N to 45°N and the meridional cross-section averaged from 40°W to 50°W (land points were excluded from the zonal mean). Dashed lines in (b) and (e) indicate the extent away from the shelf that ρ was averaged over to compute ρ_{WB} and ρ_{EB} (the contribution of ρ_{EB} in CCSM3 was negligible because $\int_H^0 \rho dz \approx 0$) in Fig. 6. White shading indicates regions that are not significant at the 95% confidence level.
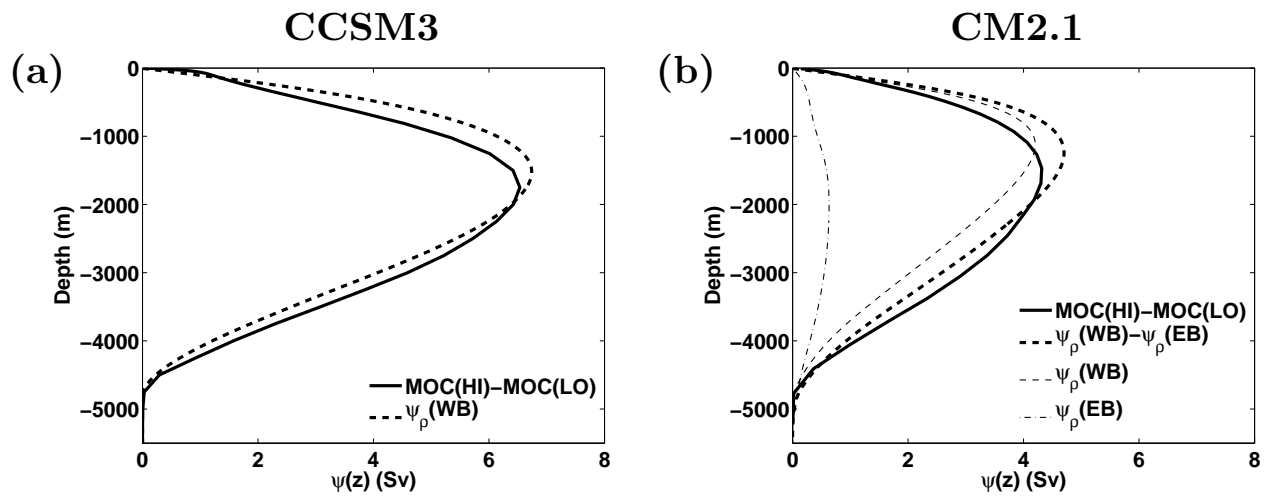
CCSM3

CM2.1

Figure 6: (HI)-(LO) index composite mean MOC averaged from $40°$N to $45°$N (solid lines) and anomalous streamfunction computed from the composite zonal cross-sections of density (dashed lines) in Figs. 5(b) and (e) using Eq. (2). Note that density anomalies on the eastern boundary are negligible in CCSM3. See text for details.
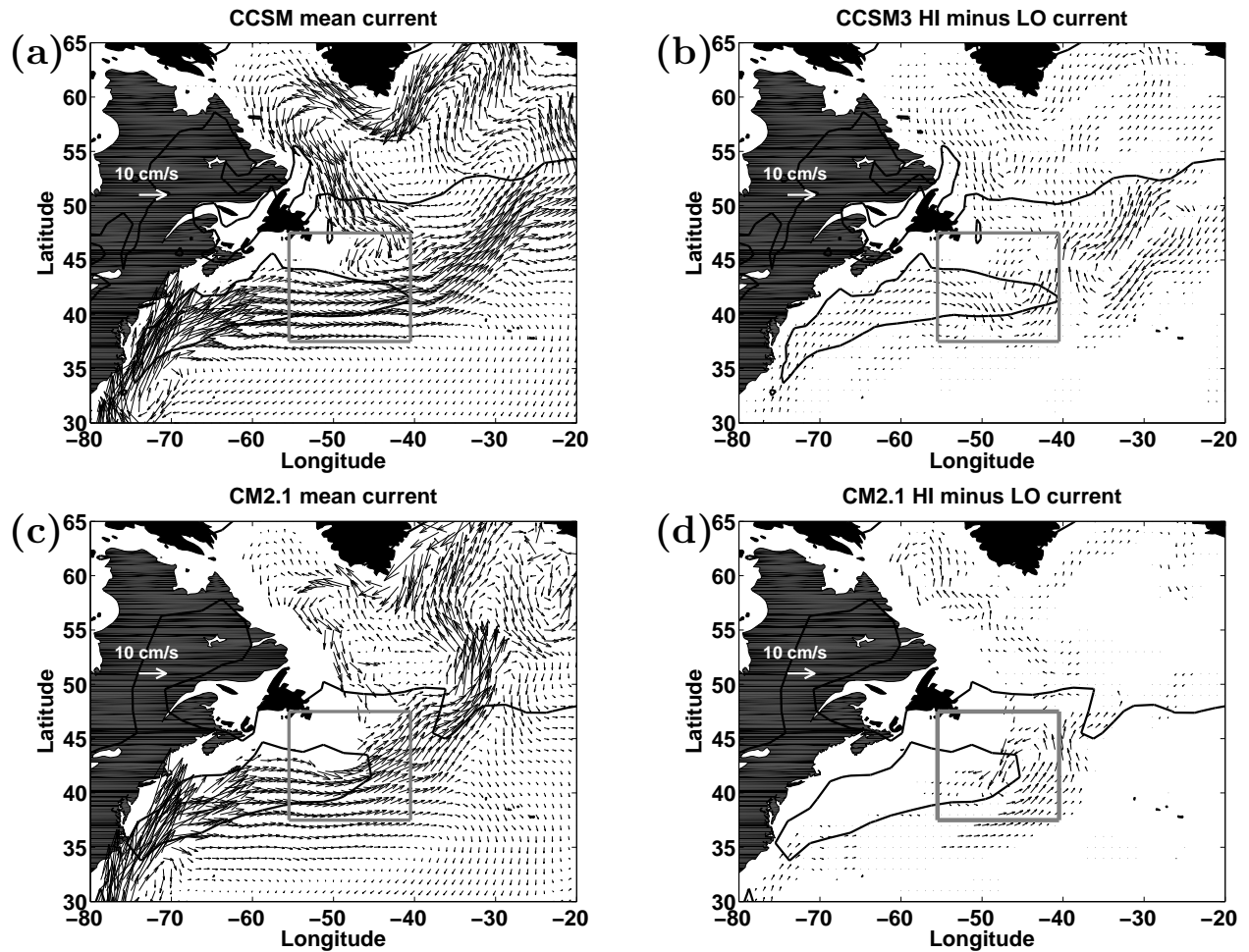
Figure 7: (a) Mean velocity and (b) composite maps of velocity using the index described in Fig. 2

near 300 m (layer 17) in CCSM3. (c) Mean velocity and (d) composites near 300 m (layer 27) in

CM2.1. Only vectors significant at the 95% confidence level are shown.
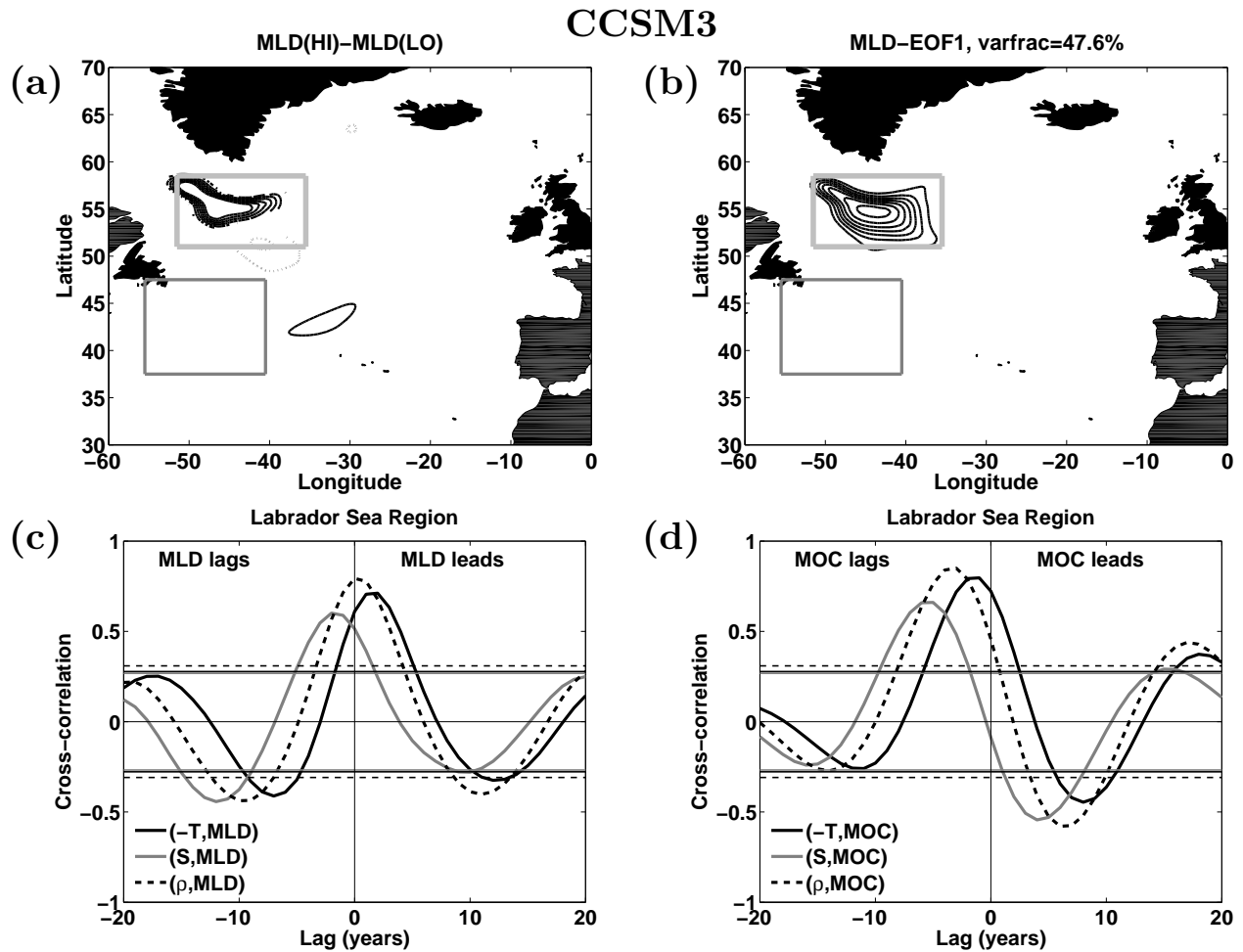
Figure 8: (a) Contours of March mixed layer depth (m) in CCSM3 using the index described in Fig. 2. Thick solid contours are 100 m to 700 m spaced by 100 m, while gray-dotted contours show -100 m and -200 m and a thin solid contour shows 0 m. (b) Contours from 100 m to 700 m of the first EOF of March mixed layer depth in CCSM3 with thick gray box indicating Labrador Sea Region for box-averaging. (c) Cross-correlation between $-T$, S and ρ against mixed layer depth in the thick boxed region shown in the Labrador Sea in (a,b). (d) Cross-correlation between $-T$, S, and ρ in the Labrador Sea Region against the MOC index defined in Fig. 1.
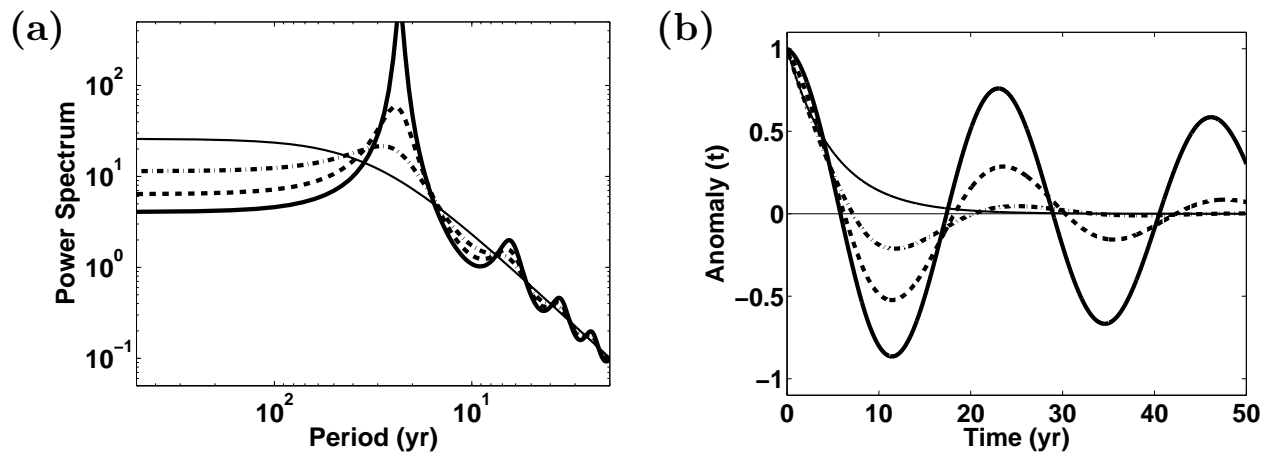
Figure 9: (a) Spectra predicted by the delayed oscillator model in Eq. (4) with parameters $\alpha = 1$, $t_d = 8$ years, $\lambda = 2\pi/32$ years^{-1}, and $D = 0$ (thin solid line), $D = 0.1$ (dash-dotted, $q \approx 1$), $D = 0.2$ (dashed, $q \approx 2.5$) and $D = 0.3$ (thick solid line, $q \approx 11$). (b) "Free solutions" to Eq. (4) as in Czaja and Marshall (2000), see text for details.
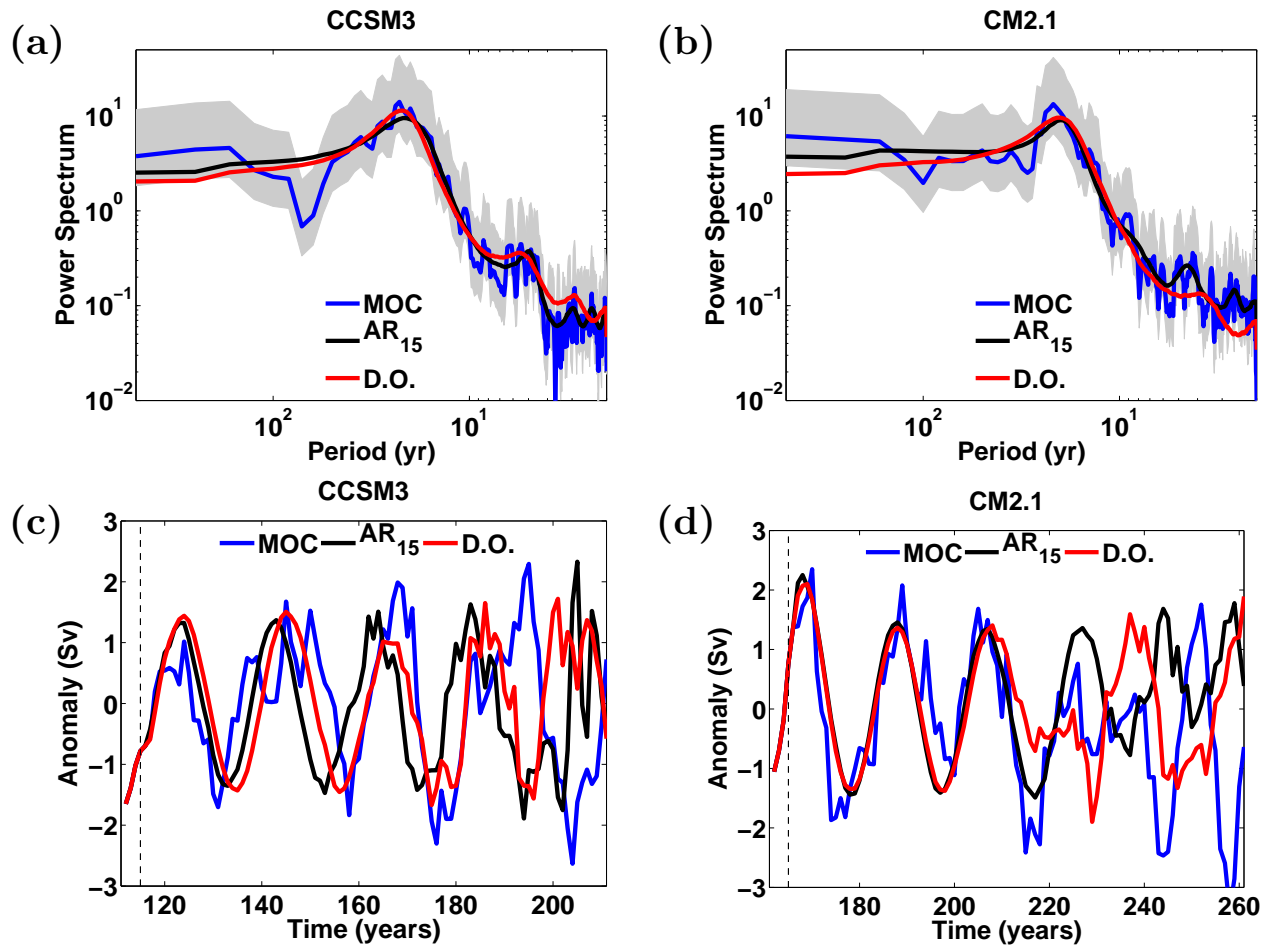
Figure 10: Fit of delayed oscillator models (red lines) to the MOC spectra (blue lines) in (a) CCSM3 and (b) CM2.1. Also plotted in black are spectra of autoregressive (AR_{15}) fits to the MOC. Example predictions for the two models are shown in (c,d).

Figure Captions

Figure1: (a) Eulerian-mean MOC (in units of Sv) in the Atlantic in CCSM3 and (b) time series of the MOC index, defined as the average MOC from $35°$N to $50°$N and 1800 m to 500 m depth, as indicated by the black box in (a). Years marked by x's (squares) denote years that are more than one standard deviation above (below) the time mean. These are used to construct composite maps. (c) The MOC averaged over years marked by x's minus the MOC averaged over years marked by squares, *i.e.,* the MOC(HI)-MOC(LO). The same variables are plotted for CM2.1 in (d), (e) and (f).

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592 left. White shading indicates regions that are not significant at the 95% confidence level.

593 **Figure 4**: Same as Fig. 3 but for CM2.1.

594 **Figure 5**: Composite zonal and meridional cross-sections of meridional velocity (m·s^{-1}) and den-

595 sity (kg m^{-3}) in CCSM3 (top row) and CM2.1 (bottom row). The zonal cross-section averaged

596 from 40°N to 45°N and the meridional cross-section averaged from 40°W to 50°W (land points

597 were excluded from the zonal mean). Dashed lines in (b) and (e) indicate the extent away from the

598 shelf that ρ was averaged over to compute ρ_{WB} and ρ_{EB} (the contribution of ρ_{EB} in CCSM3 was

599 negligible because $\int_H^0 \rho dz \approx 0$) in Fig. 6. White shading indicates regions that are not significant

600 at the 95% confidence level.

601 **Figure 6**: (HI)-(LO) index composite mean MOC averaged from $40°$N to $45°$N (solid lines) and

602 anomalous streamfunction computed from the composite zonal cross-sections of density (dashed

603 lines) in Figs. 5(b) and (e) using Eq. (2). Note that density anomalies on the eastern boundary are

604 negligible in CCSM3. See text for details.

605 **Figure 7**: (a) Mean velocity and (b) composite maps of velocity using the index described in Fig. 2

606 near 300 m (layer 17) in CCSM3. (c) Mean velocity and (d) composites near 300 m (layer 27) in

607 CM2.1. Only vectors significant at the 95% confidence level are shown.

608 **Figure 8**: (a) Contours of March mixed layer depth (m) in CCSM3 using the index described in

609 Fig. 2. Thick solid contours are 100 m to 700 m spaced by 100 m, while gray-dotted contours

610 show -100 m and -200 m and a thin solid contour shows 0 m. (b) Contours from 100 m to 700 m of

611 the first EOF of March mixed layer depth in CCSM3 with thick gray box indicating Labrador Sea

612 Region for box-averaging. (c) Cross-correlation between $-$T, S and ρ against mixed layer depth in

613 the thick boxed region shown in the Labrador Sea in (a,b). (d) Cross-correlation between $-T$, S,

614 and ρ in the Labrador Sea Region against the MOC index defined in Fig. 1.

615 **Figure 9**: (a) Spectra predicted by the delayed oscillator model in Eq. (4) with parameters $\alpha = 1$,

616 $t_d = 8$ years, $\lambda = 2\pi/32$ years^{-1}, and $D = 0$ (thin solid line), $D = 0.1$ (dash-dotted, $q \approx 1$),

617 $D = 0.2$ (dashed, $q \approx 2.5$) and $D = 0.3$ (thick solid line, $q \approx 11$). (b) "Free solutions" to Eq. (4)

618 as in Czaja and Marshall (2000), see text for details.

619 **Figure 10**: Fit of delayed oscillator models (red lines) to the MOC spectra (blue lines) in (a)

620 CCSM3 and (b) CM2.1. Also plotted in black are spectra of autoregressive (AR$_{15}$) fits to the

621 MOC. Example predictions for the two models are shown in (c,d).