

1 High frequency variability of the Atlantic meridional
2 overturning circulation

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Abstract

10 We compare the variability of the Atlantic meridional overturning circulation
11 (AMOC) as simulated by the coupled climate models of the RAPID project,
12 which cover a wide range of resolution and complexity, and observed by the
13 RAPID/MOCHA array at about 26°N . We analyse variability on a range of
14 timescales, from five-daily to interannual. In models of all resolutions there
15 is substantial variability on timescales of a few days; in most AOGCMs the
16 amplitude of the variability is of somewhat larger magnitude than that ob-
17 served by the RAPID array, while the time-mean is within about 10% of the
18 observational estimate. The amplitude of the simulated annual cycle is similar
19 to observations, but the shape of the annual cycle shows a spread among the
20 models. A dynamical decomposition shows that in the models, as in observa-
21 tions, the AMOC is predominantly geostrophic (driven by pressure and sea-level
22 gradients), with both geostrophic and Ekman contributions to variability, the
23 latter being exaggerated and the former underrepresented in models. Other
24 ageostrophic terms, neglected in the observational estimate, are small but not
25 negligible. The time-mean of the western boundary current near the latitude of
26 the RAPID/MOCHA array has a much wider model spread than the AMOC
27 does, indicating large differences among models in the simulation of the wind-
28 driven gyre circulation, and its variability is unrealistically small in the models.
29 In many RAPID models and in models of the Coupled Model Intercompari-
30 son Project Phase 3 (CMIP3), interannual variability of the maximum of the
31 AMOC wherever it lies, which is a commonly used model index, is similar to in-
32 terannual variability in the AMOC at 26°N . Annual volume and heat transport
33 timeseries at the same latitude are well-correlated within $15\text{--}45^{\circ}\text{N}$, indicating

34 the climatic importance of the AMOC. In the RAPID and CMIP3 models, we
35 show that the AMOC is correlated over considerable distances in latitude, but
36 not the whole extent of the north Atlantic; consequently interannual variability
37 of the AMOC at 50°N, where it is particularly relevant to European climate, is
38 not well-correlated with that of the AMOC at 26°N, where it is monitored by
39 the RAPID/MOCHA array.

40 1 Introduction

41 Any substantial change, whether anthropogenic or natural, in the meridional over-
42 turning circulation of the Atlantic Ocean (AMOC) could considerably affect the
43 climate, especially of the north Atlantic and Europe, on account of the associated
44 northward ocean heat transport. A complete cessation of the AMOC would produce
45 a strong cooling (Vellinga and Wood, 2002; Stouffer et al., 2006), but this is very
46 unlikely during the 21st century according to the latest assessment of the Intergov-
47 ernmental Panel on Climate Change (Meehl et al., 2007). Schmittner et al. (2005)
48 and Meehl et al. (2007) show that there exists a wide range of weakening—from
49 0% to 50%—of the AMOC by 2100 in model projections of climate change under
50 scenarios of increasing anthropogenic greenhouse gas concentrations. Other studies
51 (Knight et al., 2005; Keenlyside et al., 2008) suggest that AMOC may weaken over
52 the next decade due to unforced (natural) variability, resulting in a cooler climate
53 around the north Atlantic. The internally generated interannual variability of the
54 AMOC in coupled AOGCMs (Dong and Sutton, 2001; Collins et al., 2006) and
55 in ocean-alone GCMs (Bjastoch et al., 2008) is found to be closely linked to in-
56 terannual variations in Atlantic Ocean heat transport (AOHT). Understanding the
57 unforced interannual variability of the AMOC and AOHT is important because it
58 is the background against which any signal of climate change has to be detected.

59 Because of such considerations, the RAPID/MOCHA array (Cunningham et al.,
60 2007; Kanzow et al., 2007; Bryden et al., 2009; Kanzow et al., 2010; Johns et al.,
61 2011) was deployed at 26.5°N in the Atlantic Ocean to monitor the AMOC and

62 provide information about its variability. The array data show temporal variability
63 in the AMOC on a broad range of time scales, from interannual to daily. The latter
64 part of the AMOC variability spectrum has not been much studied in the numerical
65 models used for climate projections. The question thus arises of whether they are
66 able to represent it realistically and if so, what the physical sources of the variability
67 are.

68 The RAPID programme, which established the observational array, also includes
69 an intercomparison project of UK global climate models (the RAPID models) of
70 varying resolution and complexity. This study reports on that project and has
71 two topics. In the first topic, we use the 5-year-long RAPID/MOCHA dataset to
72 evaluate and compare the RAPID models in regard to high-frequency variability,
73 which is a new kind of observational information. In the second topic, we set the
74 high-frequency observations at 26°N into their climatic context, by analysing the
75 relationship between volume transport and heat transport at different timescales and
76 at various latitudes in the north Atlantic. The connection between these topics, and
77 the motivation for the study, is the dataset from the RAPID/MOCHA monitoring
78 array at 26°N.

79 Model intercomparison is valuable for assessing model systematic uncertainty
80 and to study its causes (e.g. Gregory et al., 2005; Stouffer et al., 2006; Griffies et
81 al., 2009). The high-frequency AMOC variability simulated by two climate models
82 is assessed in Baehr et al. (2009) using the first year of data from the RAPID array.
83 They found that the magnitude of variability is well reproduced in ECHAM5/MPI-

84 OM, and ECCO-GODAE shows significant correlation of the daily AMOC to that
85 of the RAPID/MOCHA time series. ECHAM5/MPI-OM is an AOGCM whereas
86 ECCO-GODAE is a data-assimilation product using an ocean-alone GCM. The
87 ECCO-GODAE time series is expected to correlate to that of RAPID array because
88 the model is forced by NCEP/NCAR reanalysis fluxes for the one-year analysis pe-
89 riod and prior to that the model solution is evolved using an optimised initial state
90 from many observational datasets. Our study is able to use a longer observational
91 timeseries and a wider range of models.

92 The common paradigm of the AMOC as a single, basin-scale, meridionally co-
93 herent zonally integrated circulation in the north Atlantic is challenged by recent
94 studies (Bingham et al., 2007; Willis, 2010; Lozier et al., 2010). Therefore the rep-
95 resentativeness of the transport measured at 26°N and its climatic impact on the
96 higher latitudes is a key question to be addressed. From the climate science point of
97 view, the main motivation for the RAPID monitoring array is the climatic influence
98 of the AMOC and how it might change in the future, and we depend on models for
99 information on the climatic influence of the AMOC on multiannual timescales.

100 **2 Data - models and observations**

101 **2.1 Models**

102 The RAPID-models, namely HadCM3, FAMOUS, FORTE, FRUGAL, GENIE, CHIME
103 and HiGEM, are all global coupled atmosphere-ocean models without flux adjust-

104 ments. They are all employed for investigations of climate variability and change on
105 various timescales. The specifications of their atmosphere and ocean components
106 are summarised in Tab. 1.

107 **HadCM3** (Gordon et al., 2000) is a Hadley Centre atmosphere–ocean gen-
108 eral circulation model (AOGCM) which has been used successfully for many pur-
109 poses and extensively cited, for instance in the IPCC Fourth Assessment Report.
110 **FAMOUS** (Jones et al., 2005, Smith et al., 2008) is a low-resolution version of
111 HadCM3, calibrated to replicate HadCM3 climate as closely as possible. It runs ten
112 times faster than HadCM3, making it a computationally less expensive AOGCM for
113 long-term or large ensembles of climate simulations. **HiGEM** (Shaffrey et al., 2009)
114 is a high-resolution AOGCM derived originally from the Hadley Centre AOGCM
115 HadGEM1. Compared to HadCM3, the predecessor of HadGEM1, HiGEM has new
116 atmospheric and sea-ice dynamics submodels together with substantial differences
117 in the ocean such as a linear-free surface, a 4th order advection scheme, 40 vertical
118 levels and the Gent-McWilliams mixing scheme being turned off. It has an eddy-
119 permitting ocean and allows fine spatial and temporal coupling between the ocean
120 and atmosphere. HiGEM is computationally expensive but several multi-decadal
121 runs with it have been completed. **FORTE** (Blaker et al., 2011) uses a recoded
122 version (MOMA, Webb, (1996)) of the Modular Ocean Model (MOM) (Pacanowski,
123 1990). It is similar to that of the Hadley Centre models and is at a resolution be-
124 tween the HadCM3 and FAMOUS ocean, but has a spectral atmospheric dynamics
125 submodel with higher resolution than the HadCM3 atmosphere, and simpler atmo-

126 spheric physics. **CHIME** (Megann et al., 2010) couples the atmosphere model of
127 HadCM3 with a predominantly isopycnic ocean (hybrid-coordinate ocean, HYCOM
128 (Bleck, 2002)), the only RAPID-model using such a scheme rather than horizontal
129 levels of fixed depth. **FRUGAL** (Bigg and Wadley, 2001) has an energy-moisture
130 balance advective-diffusive atmospheric component, based on the UVic model of
131 Weaver et al. (2001). It does not simulate winds, and a prescribed wind-stress
132 climatology is applied to the ocean. FRUGAL uses the MOM ocean with a grid
133 designed to improve resolution of the Arctic Ocean. **GENIE** (Edwards and Marsh,
134 2005) also uses the UVic atmosphere and is the only RAPID-model which does
135 not have a primitive-equation ocean model; instead, it uses a frictional geostrophic
136 model (GOLDSTEIN) in which horizontal momentum diffusion is parameterised by
137 Rayleigh friction rather than viscosity. This is computationally very cheap and con-
138 sequently GENIE is the fastest RAPID-model by a large factor, suiting its intended
139 use for multimillennial climate simulations and very large ensembles.

140 For this analysis, we produced 10 years of 5-daily model data (i.e. 5-day means)
141 from the unforced control integrations of the models. Control integrations are cus-
142 tomarily evaluated with respect to present-day climatology, especially for internal
143 variability. This simplifies comparison of model and observational results by avoid-
144 ing the complications of whether radiative forcings of climate change are the same
145 in different climate models and whether trends associated with climate change are
146 realistically simulated. For calculation of the interannual variability of the model
147 AMOC, we also produced time-series of 110 years of annual means from the control

148 integrations. The data analysed in this paper comes from portions of the control
149 runs after the models have been spun up for many hundred years except in HiGEM
150 and CHIME where the control runs are only 115 and 200 years long, respectively.
151 The 5-daily data in CHIME and HiGEM is from year 60 to year 70. The annual
152 data in CHIME is from year 60 to year 170, and in HiGEM from year 20 to year
153 110, only 90 years long, after a short spin-up time.

154 **2.2 Observations**

155 The RAPID/MOCHA array is the first system able to monitor a basin-wide trans-
156 port at a latitude continuously. It is designed to estimate the AMOC as the sum of
157 three observable components namely, Ekman transport, Florida Current transport
158 and the upper mid-ocean transports (See Sect. 4 for more details). Note that it is an
159 observational estimate of a composite of the main contributions with an unknown
160 residual term that is assumed to be small and barotropic. It does not include other
161 ageostrophic components than the Ekman component. The array has temporally
162 high sampling, i.e. 12 hourly but does not have spatially high sampling across the
163 latitude and depths. The observational timeseries are 5 years long, from April 2004
164 to March 2009. We average the 12-hourly measurements (10-day low-pass filtered)
165 to produce 5-daily data for comparison to the 5-daily model data. The 5-daily data
166 has a standard deviation only 3.2% less than that of the 12-hourly data.

167 3 Comparison of simulated and observed variability

We calculate the timeseries of the 5-daily Atlantic meridional overturning transport at about 26°N in models and measurements. The overturning transport T_{over} at a given latitude y and time t is the zonal and vertical integral of the meridional velocity v

$$T_{over}(y, t) = \int \int_z^0 v(x, y, z', t) dz' dx \quad (1)$$

168 where x and z are the zonal and vertical axes respectively and the zonal integral
169 is across the whole width of the Atlantic basin. We take the depth integral from
170 the surface ($z' = 0$) to a depth of $z' \simeq 1000$ m (or to the bottom at longitudes
171 where the ocean is shallower than z), to include all of the northward branch of
172 the AMOC. The precise latitude and depth for evaluating T_{over} are chosen for each
173 model to coincide with a boundary between model cells in each direction and are
174 shown in Tab. 1. By construction, the value of T_{over} is identical with the meridional
175 overturning streamfunction at the given latitude and depth. At about 26°N, all
176 models have a long-term mean strength in the range 16–21 Sv, within 10% of the
177 observed 18.6 Sv (Table 1). HiGEM has the smallest time-mean and FAMOUS the
178 largest.

179 Substantial variability on short time scales is evident in models as well as in
180 observations in the timeseries for a single year (Figure 1a), shown as an illustration.
181 Calculating the 5-daily standard deviation at 26°N for this single year gives 3–5 Sv
182 for the observations and all the models except FRUGAL and GENIE (Tab. 1).

183 This is remarkable, given the wide range of complexity of the models, and it is
184 interesting that the magnitude of simulated variability does not depend on model
185 resolution. GENIE and FRUGAL have no high-frequency variability. These models
186 use the UVic atmosphere model which does not have internal dynamics capable
187 of generating variability. In both the models, ocean is forced by prescribed annual
188 wind-stress climatology. It is likely that in the other models the atmosphere provides
189 most of the ocean variability (Gregory et al., 2005). Indeed, when the GENIE ocean
190 is coupled to a dynamical atmosphere (Lenton et al., 2007), notable interannual
191 AMOC variability is generated.

192 A single year is not representative of climatological statistics, so we calculate
193 the mean annual cycle from the 10 individual years for each model and the 5 years
194 of observations (Figure 1b). The high-frequency variability is thereby reduced, but
195 still notable; the 5-daily standard deviation remains similar across most models and
196 is slightly larger in observations (Tab. 1). Part of the variability comes from the
197 annual cycle. The observations show a maximum in autumn and a minimum in
198 spring whereas the models show a range of seasonal behaviour (Figure 2).

199 The variance spectra of the time series (Figure 1c) show that the annual cycle
200 is the dominant period in both models and observations. In all the models, its
201 variance is within a factor of two of that of observations. At the highest frequencies,
202 however, all the models except CHIME have greater variance than observations, by
203 up to an order of magnitude, with no systematic dependence on model resolution.
204 FAMOUS shows particularly large variance in shorter periods. CHIME shows least

205 variance both for the annual cycle and at high frequencies. Since it uses the same
206 atmosphere model as HadCM3, this difference must be due to the ocean model in
207 some way. Oscillations of less than 40-day period are significant in all the models
208 (except FRUGAL and GENIE) and observations.

209 The results we describe in this section and the next are based on the 5 years
210 of observations available so far and 10 years of model data. We reach the same
211 conclusions if we use either the first 5 years of the model data or the last 5 years
212 i.e. the same length as the observations, instead of ten years. The 5-daily standard
213 deviation of each year of the simulations and observations are shown in Figure 3.

214 4 Dynamical decomposition of the transport

215 In order to identify the physical sources of variability in the simulated overturning,
216 a dynamical decomposition of the transport is carried out on the 5-daily timeseries.
217 Previous modelling studies (Lee and Marotzke, 1998; Hirschi et al., 2003; Sime et al.,
218 2006; Baehr et al., 2009) suggest various ways of decomposing the transport. Cun-
219 ningham et al. (2007) obtain the observational T_{over} from Ekman, Florida Current
220 and upper mid-ocean components, of the RAPID/MOCHA array. The Ekman com-
221 ponent is physically distinguished; it exists within the upper tens of metres which are
222 affected by the windstress and the vertical shear it causes. The Florida Current com-
223 ponent is geographically distinguished; it is the integral of flow at all depths passing
224 through the narrow channel between Florida and the Bahamas, within which there
225 is a specific monitoring system. The channel is 800 m deep and the flow through

226 it is entirely counted in the northward branch of the AMOC. The upper mid-ocean
 227 component is the geostrophic meridional flow above 1100 m through the 26.5°N
 228 section across the Atlantic from the Bahamas to Africa.

229 Florida and the Bahamas are not represented with realistic geography, or at all,
 230 in the models. Hence we cannot meaningfully calculate the Florida Straits transport,
 231 and instead we carry out the decomposition slightly further north, at around 29°N,
 232 between the coasts of America and Africa. (At the end of this section, we evaluate
 233 the western boundary current in the models.) Again, the precise latitude is model-
 234 dependent, and the same depth is used as for 26°N (Tab. 1). Our decomposition
 235 of T_{over} is physically based, consistent with the model formulations, into Ekman,
 236 geostrophic, viscous and advective components.

Consider the equation of motion. The zonal acceleration is given as

$$\frac{Du}{Dt} = \mathbf{u} \cdot \nabla u + \frac{\partial u}{\partial t} = -\frac{1}{\rho} \frac{\partial P}{\partial x} + fv + F_v + F_h \quad (2)$$

where \mathbf{u} is the 3D velocity and u its eastward component, $\partial P/\partial x$ is the zonal pressure gradient, f is the Coriolis parameter, $F_v = \kappa \partial^2 u / \partial z^2$ is the vertical momentum diffusion term with κ the coefficient of vertical viscosity, $F_h = \eta_{Lap} \nabla_H^2 u$ and/or $F_h = \eta_{bi} \nabla_H^4 u$ (according to model formulation) is the horizontal momentum diffusion term with η_{Lap} and η_{bi} being the coefficients of horizontal viscosity, and ρ is the Boussinesq reference density. We rearrange Eq.(2) and integrate it over depth and

longitude across the Atlantic as

$$\int \int_z^0 v dz' dx = \frac{1}{f} \int \int_z^0 \left(\frac{1}{\rho} \frac{\partial P}{\partial x} - F_v - F_h + \mathbf{u} \cdot \nabla u + \frac{\partial u}{\partial t} \right) dz' dx \quad (3)$$

237 Thus we treat the total transport on the LHS as a sum of the terms on the RHS as
 238 follows.

239 The geostrophic transport (T_{geo}) is the term due to $\partial P/\partial x$ and consists of two
 240 parts: the internal part (T_{int}), which is due to the pressure gradient $\partial P_\rho/\partial x$ caused
 241 by zonal density gradients, and the external part (T_{ext}), which is due to the sea
 242 surface slope $\partial h/\partial x$ in models with a free surface (HiGEM, FORTE) or to the rigid
 243 lid pressure gradient $\partial P_s/\partial x$ in rigid lid models (HadCM3, FAMOUS and GENIE),
 244 where effectively $P_s = h\rho g$. Thus

$$T_{geo} = T_{ext} + T_{int}, \quad T_{int} = \frac{1}{\rho f} \int \int_z^0 \frac{\partial P_\rho}{\partial x} dz' dx, \quad T_{ext} = \frac{1}{\rho f} \int \int_z^0 \frac{\partial P_s}{\partial x} dz' dx \quad (4)$$

The vertical momentum diffusion $\kappa \partial^2 u/\partial z^2$ is the vertical derivative of the diffusive vertical momentum flux $\kappa \partial u/\partial z$. Integrated over the upper ocean, this equals the surface momentum flux i.e. the zonal wind stress τ_x , which is all absorbed in the Ekman layer. The bottom boundary layer is far below, and the bottom stress is identically zero in HadCM3 and FAMOUS, which have a free-slip bottom boundary condition, and is negligible in HiGEM and FORTE. GENIE has no bottom boundary layer or explicit bottom stress. Hence there is no contribution from bottom stress

to the Ekman transport

$$T_{Ek} = -\frac{1}{\rho f} \int \tau_x dx. \quad (5)$$

The ageostrophic transport due to the horizontal momentum diffusion i.e. horizontal viscosity is

$$T_{vis} = -\frac{1}{f} \int \int_z^0 \eta_{Lap} \nabla_H^2 u dz' dx \quad \text{and/or} \quad T_{vis} = -\frac{1}{f} \int \int_z^0 \eta_{bi} \nabla_H^4 u dz' dx \quad (6)$$

245 The horizontal diffusion terms are Laplacian ($\nabla_H^2 u$) and/or biharmonic ($\nabla_H^4 u$) for-
 246 mulations with different coefficient of viscosity in each model. In theory these viscous
 247 terms represent the horizontal momentum flux due to unresolved eddies, although in
 248 practice horizontal viscosity is increased to ensure model dynamical stability. The
 249 viscous term can locally be of either sign, since its effect is to transport momentum.
 250 Globally, it must sum to zero for momentum, but is a positive definite sink of kinetic
 251 energy.

The advective transport (T_{adv}) due to the non-linear advective term $\mathbf{u} \cdot \nabla u$ is

$$T_{adv} = \frac{1}{f} \int \int_z^0 \mathbf{u} \cdot \nabla u dz' dx \quad (7)$$

252 where the momentum flux due to resolved eddies would appear. This term is absent
 253 in GENIE by construction.

In HadCM3, FAMOUS and HiGEM we can calculate all the components. Any residual is due to acceleration $\partial u / \partial t$. The residual due to the local acceleration is

negligibly small and is ignored in all models, so

$$T_{over} = T_{geo} + T_{Ek} + T_{vis} + T_{adv} \quad (8)$$

254 As an example, this decomposition is shown for HadCM3 in Figure 4. In GENIE,
255 we calculate T_{over} , T_{Ek} and T_{vis} , and infer T_{geo} as a residual. This model uses an
256 annual climatology of windstress as a constant term, so T_{Ek} does not contribute
257 to variability. In FORTE, we calculate T_{over} , T_{Ek} and T_{vis} due to the Laplacian
258 diffusion term, and infer T_{geo} as the residual. This means that the biharmonic
259 diffusion term is included in T_{geo} . This term is implicit in the model (Webb et al.,
260 1998) and could not be calculated offline. It is relatively large and it is unclear how
261 to interpret it physically. The components of transport could not be computed for
262 FRUGAL and CHIME.

263 The mean and 5-day variability of the components of observed and simulated
264 transports are shown in Tab. 1. The observed geostrophic transport is the sum of the
265 mid-ocean transport and Florida current transport. In the mean, the geostrophic
266 term is largest in all cases. The Ekman term is relatively small and positive, and
267 the viscous term even smaller and negative, except in GENIE, in which the viscous
268 (actually frictional) term is larger than in other models and the signs of these two
269 terms are the other way round.

270 As discussed above, the largest part of the variability is the mean annual cycle.
271 The two main sources of this variability are T_{Ek} (Figure 5a) and T_{geo} (Figure 5b) in

272 the models, as in observations (Cunningham et al., 2007). However, T_{geo} variability
273 is smaller than T_{Ek} variability in models whereas in observations the reverse is true
274 (Tab. 1). It is evident in Figure 4 that the Ekman term dominates the annual cycle
275 in HadCM3, for example.

276 We find that T_{geo} variability tends to be underestimated in models as compared
277 to observations. In the observations, the variability is found to be due to the effect
278 of the seasonal momentum flux on the eastern boundary density (Chidichimo et
279 al., 2010; Kanzow et al., 2010). This suggests that models might underestimate the
280 variability of the pressure anomaly along the eastern/western boundaries, possibly as
281 the result of underestimating the adiabatic upwelling/downwelling processes driven
282 by alongshore wind-stress due to the coarse resolution which spreads the effect over
283 one grid box instead of a more confined area in reality. As the geostrophic seasonal
284 cycle is mainly driven by surface fluxes, unrealism in either the surface fluxes or the
285 vertical mixing caused by the surface fluxes could also be a cause of underestimated
286 variability in models. In eddy-permitting HiGEM, the geostrophic seasonal cycle
287 has more variability than in HadCM3 (Figure 5c), and dominates the shape of the
288 annual cycle, as in observations. This is true also of FORTE, but in that case the
289 “geostrophic” term actually includes a large residual due to the biharmonic diffusion
290 (as noted above).

291 As in the observed variability (Kanzow et al., 2007), the external T_{ext} and internal
292 T_{int} components of T_{geo} in the upper 1000 m strongly anticorrelate in most models
293 (Tab. 1) since by construction, $T_{geo}(z, t) = T_{int}(z, t) + T_{ext}(z, t)$, where z is a suitably

294 chosen depth, so that $dT_{int}/dt = -dT_{ext}/dt + dT_{geo}/dt$. Indeed, this expression shows
 295 that a strong anticorrelation between T_{int} and T_{ext} should be observed whenever the
 296 fluctuations in T_{geo} become small relative to that of T_{ext} and T_{int} , mathematically
 297 when $|dT_{geo}/dt| \ll |dT_{int}/dt|$, which when it occurs expresses deep compensation.
 298 According to classical theories describing the spin-up of a stratified ocean in response
 299 to change in wind forcing, e.g., Anderson and Killworth (1977), Anderson and Corry
 300 (1985), the physical mechanism for such a deep compensation is speculated to be
 301 associated with the baroclinic adjustment by oceanic Rossby waves, which is usually
 302 found to compensate the barotropic response (that usually characterizes the initial
 303 stages of the adjustment to a change in the wind forcing) in the deeper layers. Note
 304 that an external component, T_{ext} , is not considered in Cunningham et al. (2007)
 305 and Kanzow et al. (2010); instead the compensation term for the mass-conservation
 306 plays this role, in effect.

307 Variability due to the viscous term T_{vis} is small but not quite negligible. This
 308 term is not calculated for the observational array, because it represents the effect
 309 of unresolved motion and, by definition, any quantity measured by the array has
 310 been “resolved” by it. The analogue of this term would be any contribution to T_{over}
 311 from ageostrophic motion; the observational estimate assumes that the motion is
 312 geostrophic or Ekman, as it has to do because the current is not directly measured
 313 at all, except in the Florida Straits and near the western boundary. Consequently
 314 the array cannot measure the ageostrophic contribution due to the advective term,
 315 which is found to be negligible in HadCM3, FAMOUS and FORTE. However, in

316 eddy-permitting HiGEM, T_{adv} makes a considerable contribution, of about 2% of the
 317 total mean transport and 17% of the total transport variability. It might therefore
 318 be a significant omission from the monitoring system.

319 Our physical decomposition does not include an explicit Gulf Stream component,
 320 which in reality passes through the Florida Straits. As discussed above, this is not
 321 geographically resolved in all the models, but we can estimate the northward western
 322 boundary current transport (T_{GS}) in the models, defined geographically. To be
 323 consistent with the latitude of our decomposition and to quantify its contribution to
 324 the geostrophic transport variability, the T_{GS} estimate is also done at about 29°N.

The T_{GS} at a given latitude y and time t is the zonal and vertical integral of
 the meridional velocity v between the western boundary, xw , and longitude, xe , and
 between the surface and z , the depth of the maximum of AMOC at about 29°N.
 The exact depth and latitude for each model are the same as stated in Tab. 1.

$$T_{GS}(y, t) = \int_{xw}^{xe} \int_z^0 v(x', y, z', t) dz' dx' \quad (9)$$

325 The eastern bound, xe , is chosen for each model separately as the longitude which
 326 gives the maximum T_{GS} in the long-term mean.

327 The T_{GS} component in all the RAPID-models are shown in Figure 6. HadCM3
 328 and FRUGAL overestimate the time-mean T_{GS} while all other models underestimate
 329 (Tab. 1). There is a much wider model spread in T_{GS} than in T_{over} , pointing to large
 330 differences in the simulations of the wind-driven gyre circulation. While the observed

331 variability is 3 Sv, the simulated variability is mostly in the 1–2 Sv range except for
332 HadCM3 with the greatest value and GENIE the least. Apart from CHIME and
333 GENIE, most models show minimum transport in autumn. The seasonal cycle of
334 the Florida Straits transport using longer observations (Atkinson et al., 2010) shows
335 a summer maximum and a winter minimum. The observed seasonal cycle using the
336 monthly means of first 4 years of RAPID/MOCHA observations is also shown in
337 Kanzow et al. (2010).

338 5 Meridional coherence of transport and its components

339 The canonical picture of a meridionally coherent overturning transport is contra-
340 dicted by recent studies such as Bingham et al. (2007), Willis, (2010) and Lozier
341 et al. (2010). Bingham et al. (2007) found in two different ocean GCMs that the
342 AMOC variability south of 40°N is dominated by high-frequency variability whereas
343 north of 40°N it is dominated by decadal variability. Based on satellite and float
344 observations of sea surface height, temperature, salinity and velocity, Willis (2010)
345 estimated the AMOC at 41°N which has smaller seasonal and interannual variabil-
346 ity than at lower latitudes. Using both hydrographic observations and a numerical
347 model, Lozier et al. (2010) detected gyre-specific decadal changes in the AMOC.

348 In Figure 7 we show the annual timeseries of T_{over} at 26°N. The observed time-
349 series is not yet long enough to assess variability on multiannual timescales. FA-
350 MOUS and CHIME have greater long-period variability than other models.

351 A commonly used AMOC index from AOGCM results is M_{max} , the maximum of

352 the overturning streamfunction, wherever it occurs, within a range of latitude and
353 depth in the Atlantic, rather than at fixed latitude and depth. The RAPID/MOCHA
354 array is intended to monitor the AMOC, by measuring the circulation at only one
355 latitude. In the model results we can investigate how well M_{\max} and T_{over} at 26°N
356 represent T_{over} at other latitudes, in order to test the conventional assumption that
357 the temporal variability of the circulation is coherent throughout the basin. GENIE
358 is omitted from this analysis because it has no high-frequency or interannual vari-
359 ability, and CHIME and FRUGAL because all required timeseries are not available.

360 Calculated from 5-day means in the RAPID-models, the time-mean M_{\max} is
361 larger than the transport at 26°N , as it must be by construction, but the variability
362 of M_{\max} is generally less (Tab. 1). In annual means, however, the two timeseries
363 have similar standard deviations. We have evaluated the same statistics from the
364 AOGCMs of the Coupled Model Intercomparison Project Phase 3 (CMIP3), finding
365 that in 16 out of 20 of them the annual standard deviation is similar in M_{\max}
366 and at 26°N (Tab. 2) (“similar” when the difference between 2 standard deviations
367 is less than 0.5 Sv); the exceptions are GISS-ER, GISS-AOM, INM-CM3.0 and
368 IAP-FGOALS1.0g. That suggests greater coherence across latitudes at longer time
369 periods. However, only ten of the CMIP3 models and three of the RAPID-models
370 have high correlation (exceeding 0.5) between the two timeseries. This is likely to
371 be because there is a time lag between 26°N and the latitudes of M_{\max} . Figure 8a
372 shows the annual standard deviation of total transport as a function of latitude. No
373 model has a well-defined maximum, but there is generally more variability in the

374 tropics, diminishing towards higher latitudes. This low-latitude variability found in
375 the AMOC and also in the AOHT is wind-induced (Klinger and Marotzke, 2000;
376 Jayne and Marotzke, 2001; Marsh et al., 2009). In a 1000-yr-long GFDL-CM2.1
377 control integration (Zhang, 2010), the maximum of interannual variability is found
378 at about 35°N.

379 Next, we calculate the temporal correlation between different latitudes of time-
380 series of annual and 5-daily volume transports and their Ekman and geostrophic
381 components, in HadCM3, FAMOUS, FORTE and HiGEM. Positive correlations are
382 found between neighbouring latitudes in all timeseries, diminishing with increas-
383 ing separation (eg., for annual timeseries in HiGEM, Figure 9). Anticorrelation is
384 found for widely spaced latitudes in the Ekman component. Since this component
385 is wind-forced, the anticorrelation must indicate opposing signs of zonal windstress,
386 occurring on opposite sides of the anomalies in atmospheric pressure and circulation
387 that produce the windstress anomalies, in particular associated with the moving
388 front between subpolar and subtropical gyres. It is notable that the anticorrelation
389 is found for both 5-daily (figure not shown) and annual data, even more pronounced
390 in the former.

391 We define the “correlation length” as a function of latitude y to be the width
392 of the range of latitudes whose timeseries have a temporal correlation exceeding 0.5
393 with the timeseries at latitude y . Within 15–60°N, the correlation lengths are typi-
394 cally 20–40° in the annual timeseries (see Tab. 1 for 26°N and Figure 9 for HiGEM).
395 Correlation lengths are greater for the annual total and the geostrophic components

396 than for the Ekman. They are also greater for annual total transports than for 5-
397 daily total transports, due to the greater coherence of the annual geostrophic com-
398 ponent. Shaffrey and Sutton (2004, their Figure 1d) and Bingham et al. (2007, their
399 Figure 2) also showed long-range coherence of annual total transport for HadCM3
400 and OCCAM models. The lowest correlation length is found at about 40°N.

401 Given the typical correlation length, we conclude that the transport measured
402 by the RAPID/MOCHA array is likely to have a correlation of less than 0.5 with
403 the AMOC strength in the mid-to-high latitude Atlantic, where it has its greatest
404 importance to climate variability (See Sect. 6). In the CMIP3 data, we test this by
405 correlating timeseries of T_{over} at 26°N and 50°N; only two models have a coefficient
406 exceeding 0.5. Correlation is increased somewhat by including lags of a few years,
407 but still does not exceed 0.5 in most cases. In models where there is a lag, vari-
408 ability of T_{over} at 50°N precedes 26°N, indicating that the forcing of the large-scale
409 geostrophic variability comes from the north. A similar relation between AMOC at
410 26°N and 50°N with a time lag of 4 years is found in GFDL-CM2.1 (Zhang, 2010).
411 The mechanism behind this time lag is caused by changes in deep water forma-
412 tion occurring at the high latitudes and initiating Kelvin waves, which propagate
413 southward along the western boundary. These coastally trapped Kelvin waves are
414 manifest as transport anomalies at each latitude as they propagate from the north
415 to the equator, eastward along the equator to the eastern boundary, and then pole-
416 ward along the eastern boundaries (Johnson and Marshall, 2002). Recently, Zhang
417 (2010), using a coupled AOGCM which represents the interior pathways of North

418 Atlantic Deep Water in the mid-latitudes as observed by Bower et al. (2009), found
419 that AMOC variations propagate in an advective manner in the mid-latitudes and
420 at the speed of Kelvin waves in the sub-tropics along the western boundary.

421 **6 Relation of northward volume transport to heat trans-** 422 **port**

423 The climatic relevance of the AMOC arises from its association with the northward
424 heat transport. The seasonal to interannual meridional Atlantic Ocean heat trans-
425 port (AOHT) variability in tropics and subtropics is associated with the wind-driven
426 Ekman transport (Klinger and Marotzke, 2000; Jayne and Marotzke, 2001; Marsh et
427 al., 2009). We assess the relationship between AMOC and AOHT by correlating the
428 annual-mean time series of the AMOC to that of the AOHT at different latitudes
429 (Figure 10) in the north Atlantic. This analysis can only be done for HadCM3,
430 FAMOUS, FORTE, HiGEM and partly for CHIME. (AOHT is unavailable for other
431 RAPID models and most of the CMIP3 models.) As expected, the time-mean heat
432 transport is maximum around 10-30°N, where it is about 1 PW (Figure 11a, Tab. 1)
433 in models. Compared to the observational estimate of Ganachaud and Wunsch
434 (2003), HiGEM and FORTE values are within the error bars of 2 of the 3 north At-
435 lantic latitudes, while HadCM3 and CHIME are closer to the estimate around 50°N.
436 FAMOUS heat transports are generally underestimated. Like T_{over} , the AOHT does
437 not have a well-defined maximum in variability as a function of latitude (Figure 11b).

438 At 35°S in the Atlantic, Dong et al. (2009) found that much of the observed north-
439 ward heat transport variability is associated with the overturning component and
440 the two are significantly correlated. Johns et al. (2011) estimated that half of the
441 array-AOHT variability at 26°N is due to the Ekman component and the other half
442 by the geostrophic component.

443 Though the volume and heat transport variations in the RAPID-models do not
444 have a similar zonal profile, in general a good degree of temporal correlation is
445 found between them at all latitudes from 15°N to 45°N (Figure 10, Figure 8b,
446 Tab. 1 for 26°N). Towards higher latitudes, the contribution due to the overturning
447 decreases. The slopes of the regression are fairly similar between 26-45°N, indicating
448 the positive volume-heat transport relationship at these latitudes. However, since
449 the AMOC at 26°N and 50°N are not strongly correlated (Section 5), we expect that
450 AOHT at 50°N, in the latitudes of the northern Europe, is not strongly correlated
451 with the AMOC at 26°N. Indeed this is the case in HadCM3, FAMOUS, FORTE,
452 CHIME and HiGEM (Tab. 1). The high-latitude AMOC index is more important
453 for climate variability because it is supposed to reflect most directly the rate of deep
454 water formation; this is obscured by wind-driven variability in the AMOC at 26°N.

455 **7 Summary and Discussion**

456 The RAPID/MOCHA array has produced a dataset which permits us to assess
457 model simulations of the AMOC in new ways. We have shown that the 5-daily
458 standard deviation of the AMOC at about 26°N simulated in the RAPID set of

459 coupled climate models is comparable to that of the RAPID/MOCHA observational
460 estimate. This is an evaluation of a property that is unlikely to have been “tuned”
461 during model development, because the observational estimate is new and recent,
462 unlike the time-mean of the AMOC, which is customarily evaluated in models. The
463 standard deviation has contributions from high-frequency variability (timescale of a
464 few days), the annual cycle and interannual variability. The models generally have
465 more high-frequency variability than that estimated from observations, and a similar
466 amplitude of annual cycle, but a spread in simulating the shape of the cycle.

467 Surprisingly, there is no systematic relation between the model resolution and
468 the magnitude of variability. This contradicts to the general assumption that if
469 the resolution is increased, variability in all time-scales will be increased. Wunsch
470 (2008) contended that eddies could possibly dominate the variability of the mea-
471 sured transport, and thereby prevent the detection of a possible trend in too short
472 records, but since recent studies such as Kanzow et al. (2009), it has been increas-
473 ingly appreciated that eddies would be swept away as coastally-trapped waves upon
474 reaching the western boundaries, leaving only a weak signal in the zonally-integrated
475 volume transport. All the models used in our study are of coarse resolution, except
476 for HiGEM, which is eddy-permitting. The relative insensitivity to model resolu-
477 tion could therefore be due to the fact that none of the models are able to generate
478 enough eddy variability for this to affect the simulated transport variability substan-
479 tially. In experiments done with different resolutions of OCCAM OGCM, it is found
480 that the eddy-resolving version produced realistic AMOC variability compared to

481 observations (Marsh et al., 2009; Cunningham and Marsh, 2010).

482 We have dynamically decomposed the variability at about 29°N (slightly north
483 of the RAPID/MOCHA array in order to avoid complications with model coast-
484 lines) into Ekman, geostrophic (i.e. due to pressure and sea-level gradient) and
485 viscous/frictional components. The AMOC at 29°N is predominantly geostrophic,
486 but the Ekman term also contributes to variability. Ekman variability is more im-
487 portant in models than in observations. Other ageostrophic terms are neglected
488 in the observational estimate, but are not negligible in models; in particular, the
489 advection of momentum makes a significant contribution to AMOC variability in
490 HiGEM. Our decomposition into the terms of the model equation of motion gives
491 information about the realism of the simulation of the relevant processes, and we
492 suggest that such a decomposition of the transport would be useful to carry out with
493 other AOGCMs. We have also quantified the western boundary current transport
494 at 29°N, for comparison with the observed Florida Straits transport. The models
495 diverge much further from the observational estimate in the time-mean of the west-
496 ern boundary current than they do with the AMOC, suggesting large differences in
497 the simulation of the wind-driven gyre. As with the geostrophic contribution to the
498 AMOC, the variability of the western boundary current is less in the models than
499 observed.

500 Though we have not narrowed down the specific mechanisms responsible for the
501 simulated high-frequency variability, our results point out the role of atmosphere in
502 setting it. In models with simple atmospheres, there is little high-frequency variabil-

503 ity.

504 In the RAPID models and in most CMIP3 AOGCMs, the magnitude of inter-
505 annual variability in the AMOC at 26°N and in the maximum of the AMOC are
506 similar, the latter being a commonly used model index. (The observational dataset
507 as yet is not long enough to assess simulated interannual variability.) We find that
508 interannual variations in Atlantic ocean heat transport are fairly well correlated at
509 each latitude with the AMOC, confirming its climatic significance and the robust-
510 ness of this relationship in models. Correlation between different latitudes is fairly
511 long-range, but does not extend over the whole basin (also found by Lozier et al.,
512 2010). Consequently the AMOC at 26°N does not have a high correlation with
513 the AMOC or with heat transport at mid-to-high latitudes. Since the latter has a
514 practical importance, and because this analysis, Zhang (2010) and Hodson and Sut-
515 ton (2011) all suggest that AMOC variability on multiannual timescales propagates
516 from north to south, it would be useful to monitor the AMOC and AOHT at higher
517 latitudes as well as the latitude of 26°N occupied by the RAPID/MOCHA array.

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Model	HadCM3	FAMOUS	FRUGAL	FORTE	GENIE	CHIME	HiGEM	OBS
Atmos res: lon x lat x level	3.75 x 2.5x 19	HadCM3 at 7.5 x 3.75 x 11	Enhanced UVic	IGCM3 T42 x 15	UVic 2D	HadCM3 at-mos x 0.83 x 38	HadGEMI at 1.25 x 0.83 x 38	
Ocean res: lon x lat x level	1.25 x 1.25 x 20	HadCM3 at 3.75 x 2.5 x 20	MOM V2 with high-res Arctic	MOM 2 x 2 x 15	GOLDSTEIN 10 x 5 x 8	HYCOM at 1.25x 1.25x25	HadGEMI at 0.33 x 0.33 x 40	
T_{over} (Sv)								
Latitude ^o N/Depth(m)	26.3/995	26.3/995		26.0/1365	26.4/1158	26.3/1050	26.9/959	26.5/1041
5-daily, 1 yr	18.8 (4.3)	19.0 (4.2)	25.9 (1.2)	16.4 (4.2)	16.4 (0.3)	15.4 (3.3)	15.1 (2.6)	19.5 (5.3)
5-daily, 10 yr	17.1 (4.1)	18.2 (4.2)	26.4 (1.4)	17.2 (4.5)	16.4 (0.3)	15.0 (3.3)	15.5 (4.0)	18.6 (4.5)
annual	16.8 (0.9)	20.6 (1.3)		17.6 (1.1)	16.5 (0)	18.8 (1.2)	16.4 (1.0)	
M_{max} (Sv)								
5d-10yrs	21.9 (2.4)	18.7 (3.0)	26.5 (1.3)	21.3 (2.5)	18.5 (0.3)		20.6 (2.5)	
annual	18.9 (0.7)	20.0 (1.3)		19.8 (1.1)	18.6 (0)	20.1 (1.7)	18.9 (1.1)	
Dynamical decomposition of T_{over} (Sv) for 5-daily means (except time-step for GENIE and geographical estimate for T_{GS})								
Latitude ^o N/Depth(m)	28.8/995	28.8/995		30/1365	30/1158		28.9/959	26.5/1041
Overturing T_{over}	18.0 (4.3)	18.1 (3.7)		16.5 (3.9)	16.1 (0.1)		15.7 (3.6)	18.6 (4.5)
Ekman T_{Ek}	0.9 (4.0)	3.5 (3.5)		1.4 (3.8)	-2.3 (0)		1.6 (3.3)	3.6 (3.2)
Geostrophic T_{geo}	17.6 (2.3)	15.3 (1.6)		15.2 (2.8)	16.8 (0.1)		14.4 (2.6)	15.0 (3.5)
Viscous T_{vis}	-0.4 (0.1)	-0.8 (0.2)		-0.1(0.1)	1.7 (0.0)		0.0 (0.0),-0.1 (0.1)	
Advective T_{adv}							0.3 (0.6)	
Correlation(T_{int}, T_{ext})	-0.98	-0.94		-0.64			-0.96	-0.83
Gulf Stream T_{GS}	43.5 (4.1)	21.2 (1.4)	48.1 (2.2)	16.9 (1.4)	22.1 (0.14)	13.2 (2.1)	16.7 (1.7)	31.9 (3.0)
Latitudinal variation of annual volume and heat transport								
Corr. length (^o lat),26 ^o N	40	24		25			28	
Latitude of M_{max} (^o N)	35-45	31-34		30-40	46-51	23-60	34-45	
Corr(T_{over} ,26 ^o N, M_{max})	0.38	0.96		0.70	0.93	0.53	0.74	
Mean AOHT,26 ^o N(PW)	1.0	0.8		1.1			1.1	
Corr(T_{over} ,AOHT),26 ^o N	0.8	0.8		0.9			0.9	
Corr(T_{over} ,26 ^o N,AOHT50 ^o N)	0.00	0.24		0.39		0.42	0.36	

Table 1: Specifications of the RAPID-models; time-mean and standard deviations (X(Y) indicates X is mean and Y is SD) of simulated Atlantic ocean meridional overturning transport (in Sv), T_{over} , at 26^oN and of the maximum of Atlantic MOC, M_{max} on 5-daily and annual timescales; time-mean and standard deviation (SD) of the simulated 5-daily T_{over} ,29^oN and its decomposed components (T_{Ek} :Ekman part, T_{geo} :geostrophic part, T_{vis} :viscous/frictional part and T_{adv} : advection part) ; time-mean of simulated annual ocean meridional heat transport (AOHT in PW),26^oN and the interannual correlation T_{over} at 26^oN with M_{max} , AOHT at 26^oN and AOHT at 50^oN. The RAPID/MOCHA observational estimate (of 5 years) is given in the last column. The observed geostrophic transport is the sum of the mid-ocean transport and Florida current transport. The 1-yr statistics given for the 5-daily T_{over} , at 26^oN, is for the second year of the model integrations and the observations. In HiGEM and FORTE, the transport component due to viscous part has 2 parts namely, by the Laplacian and biharmonic terms. In FORTE, the biharmonic term is implicit and could not be calculated offline. The FRUGAL transport at 26^oN is calculated along a curvilinear gridline which is near 26^oN. Time-step data is used in GENIE which has an ocean time-step of 3.65 days. GENIE and FRUGAL have no seasonal variability in wind-stress and no interannual variability. The Gulf Stream component (T_{GS}) is not part of the physical decomposition; it is estimated geographically (See Sect. 4 for details). Meridional correlation length (in ^olat) at 26^oN is defined as the latitudinal extent of positive correlation above 0.5 in both directions. FRUGAL and CHIME data are only available for some of the calculations.

Model	SD M_{\max}	SD $T_{\text{over}26^\circ\text{N}}$	Corr ($T_{\text{over}26^\circ\text{N}}$, M_{\max})	SD $T_{\text{over}50^\circ\text{N}}$	Corr ($T_{\text{over}26^\circ\text{N}}$, $T_{\text{over}50^\circ\text{N}}$)	Lag (years)	Lagged Corr. ($T_{\text{over}26^\circ\text{N}}$, $T_{\text{over}50^\circ\text{N}}$)
CSIRO-Mk3.0	1.8	1.6	0.85	1.6	0.53	-1	0.70
CNRM-CM3	1.8	2.1	0.20	1.7	0.05	-2	0.41
CCCMA-CGCM3.1(T63)	0.72	0.71	0.85	0.67	0.11	-1	0.51
CCCMA-CGCM3.1(T47)	0.50	0.63	0.09	0.65	-0.14	-2	0.39
BCCR-BCM2-0	0.93	0.91	0.61	0.82	-0.02	-2	0.25
GISS-ER	2.7	0.97	0.06	2	0.35	-1	0.48
GISS-AOM	7.2	1.5	0.01	2.0	0.19	-3	0.44
GFDL-CM2.1	1.3	1.2	0.39	1.1	-0.01	-5	0.46
GFDL-CM2.0	1.1	1.1	0.38	1.1	0.12	-2	0.51
CSIRO-Mk3.5	1.2	1.0	0.88	1.4	0.52	-1	0.72
MIROC3.2(hires)	0.8	1.0	0.16	0.82	0.02	-1	0.28
INM-CM3.0	2.9	3.4	0.47	1.7	0.07	-2	0.52
INGV-ECHAM4	1.6	1.9	0.61	1.5	0.09	-3	0.58
IAP-FGOALS1.0g	2.3	0.49	0.09	0.43	-0.26	10	-0.02
NCAR-CCSM3.0	1.8	1.2	0.88	1.1	0.24	-2	0.45
MRI-CGCM2.3.2a	0.71	0.73	0.53	0.97	-0.23	-1	0.34
MIUB-ECHOG	1.3	1.0	0.35	1.2	0.23	-4	0.53
MIROC3.2(medres)	0.72	0.64	0.67	0.69	0.07	-2	0.44
UKMO-HadGEM1	1.0	1.0	0.68	0.77	0.05	-1	0.21
UKMO-HadCM3	1.7	1.8	0.54	1.2	0.05	1	0.21

Table 2: Comparison of standard deviations (in Sv) of Atlantic MOC (T_{over}) at 26°N , 50°N and of the maximum of Atlantic MOC, M_{\max} , and their correlations in the CMIP3 models. Linear or quadratic trend is removed for unsteady runs before the calculation. The lag between T_{over} at 26°N and 50°N is shown which gives the largest correlation of their timeseries. The lag is negative when $T_{\text{over}26^\circ\text{N}}$ lags.

719 FIGURE CAPTIONS

720 Figure 1: Atlantic MOC (T_{over}) at 26°N a) 5-daily time series - for a single year (the
721 second year of the model integrations and observations) b) 5-daily time series
722 - 10-year mean in models and 5-year mean in observations (The FRUGAL
723 transport is calculated along a curvilinear gridline which is near 26°N. For
724 GENIE, time-step data is plotted ; its ocean time-step is 3.65 days) and c) 5-
725 daily - power spectrum (Note the logarithmic scale on the y-axis. Oscillations
726 of less than 40-day period are significant in observations and in all the models,
727 except FRUGAL and GENIE).

728 Figure 2: 5-yr timeseries of 5-daily Atlantic MOC (T_{over}) at 26°N in observations
729 and in the RAPID-AOGCMs. (Data with a 45-day moving average is shown in
730 blue.) Other RAPID-models, GENIE and FRUGAL with simple atmospheric
731 components, have little interannual variability. The last 5 years of the 10 years
732 of data from each AOGCM is shown here.

733 Figure 3: Standard deviation of the 5-daily Atlantic MOC (T_{over}) at 26°N for
734 each year of simulations and observations. (Since the observational timeseries
735 starts in April, the SD is calculated from April to March and some models are
736 missing a year because of wanting to start all the years in April.)

737 Figure 4: Decomposition of 5-daily Atlantic MOC (T_{over}) into physical compo-
738 nents at about 29°N in HadCM3. The sum E+g+vis (dash-dotted) is almost
739 coincident with the total overturning (solid).

740 Figure 5: Annual cycle of 5-daily Atlantic MOC (T_{over}) components at about 29°N

741 - a) Ekman component (T_{Ek}) and b) Geostrophic component (T_{geo}).

742 Figure 6: Annual cycle of 5-daily Western Boundary Current (T_{GS}) at about 29°N

743 calculated geographically (See Sect. 4 for details).

744 Figure 7: Annual time series of the Atlantic MOC (T_{over}) at 26°N (HiGEM data

745 is only 90 years long after the spin-up time).

746 Figure 8: Zonal profile of a) annual ocean meridional overturning transport (T_{over})

747 variability (Sv) and b) correlation of annual T_{over} and ocean meridional heat

748 transport in the north Atlantic.

749 Figure 9: Cross-correlation of ocean meridional overturning transport, T_{over} and

750 its physical components, between latitudes in the north Atlantic in HiGEM:

751 Annual T_{over} (top left), geostrophic, T_{geo} (top right), Ekman, T_{ek} (bottom

752 left) and their meridional correlation length (bottom right). Correlation length

753 (°lat) as a function of latitude y is defined as the width of the range of latitudes

754 whose timeseries which have a temporal correlation exceeding 0.5 with the

755 timeseries at latitude y .

756 Figure 10: Scatter plot of annual-mean ocean meridional overturning transport,

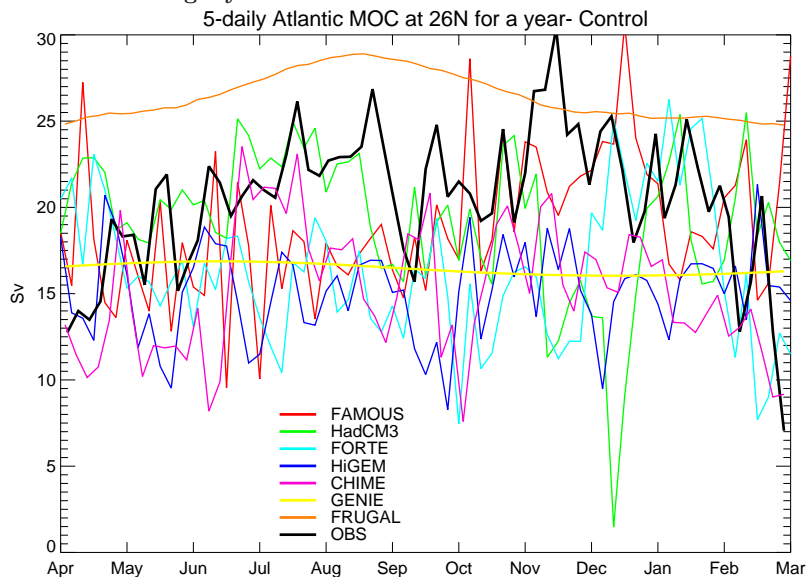
757 T_{over} (Sv) and ocean meridional heat transport (PW) at various latitudes in

758 the north Atlantic in different models. The correlation coefficients and slopes

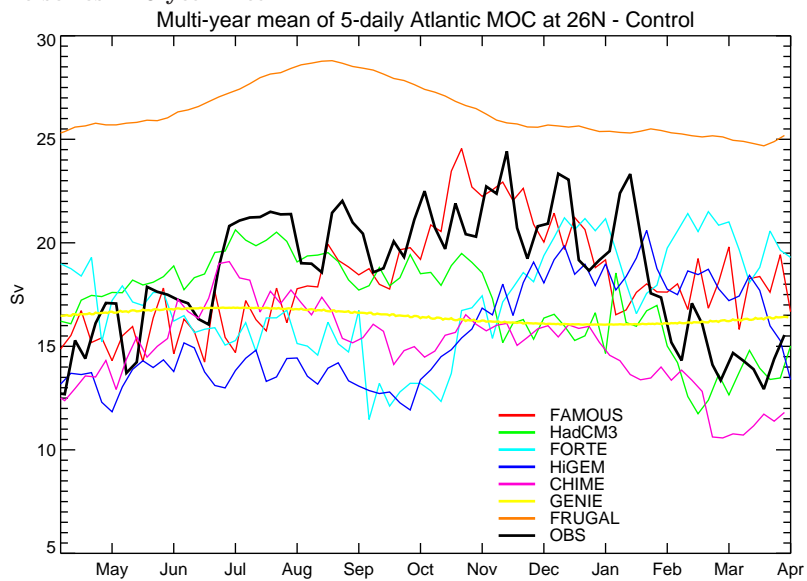
759 of the regression are given in brackets.

760 Figure 11: Zonal profile of a) mean annual ocean meridional heat transport (PW)
761 and b) variability of annual ocean meridional heat transport in the north
762 Atlantic. The observational estimate of heat transport is from Ganachaud
763 and Wunsch (2003). CHIME data is only available in 10° latitude intervals.

a) 5-daily time series - a single year



b) 5-daily time series - 10-year mean



c) 5-daily - power spectrum from 5 years of data

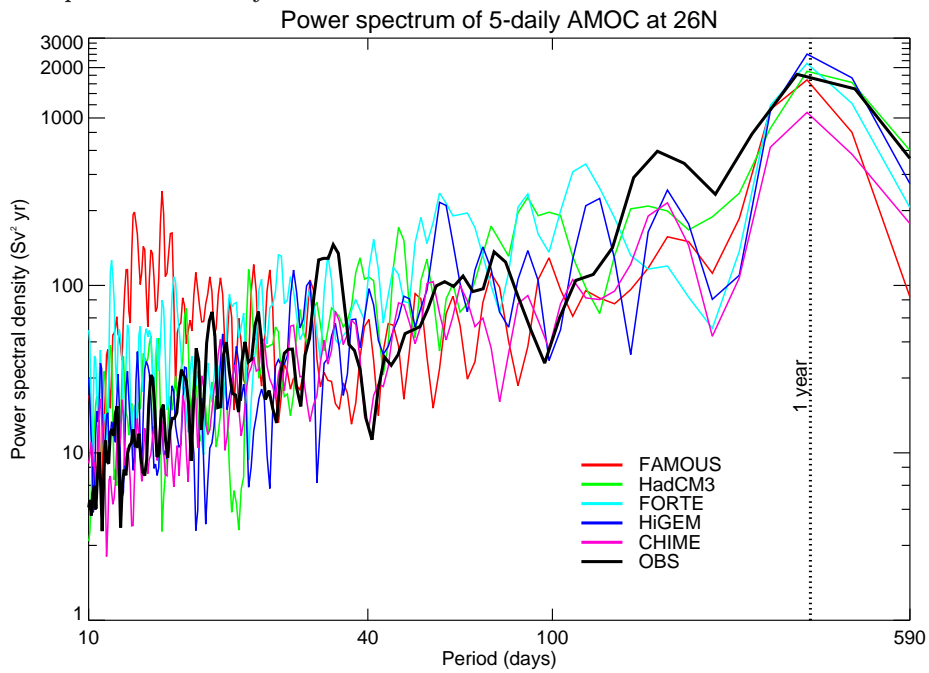


Figure 1: Atlantic MOC (T_{over}) at 26°N a) 5-daily time series - for a single year (the second year of the model integrations and observations) b) 5-daily time series - 10-year mean in models and 5-year mean in observations (The FRUGAL transport is calculated along a curvilinear gridline which is near 26°N. For GENIE, time-step data is plotted ; its ocean time-step is 3.65 days) and c) 5-daily - power spectrum (Note the logarithmic scale on the y-axis. Oscillations of less than 40-day period are significant in observations and in all the models, except FRUGAL and GENIE).

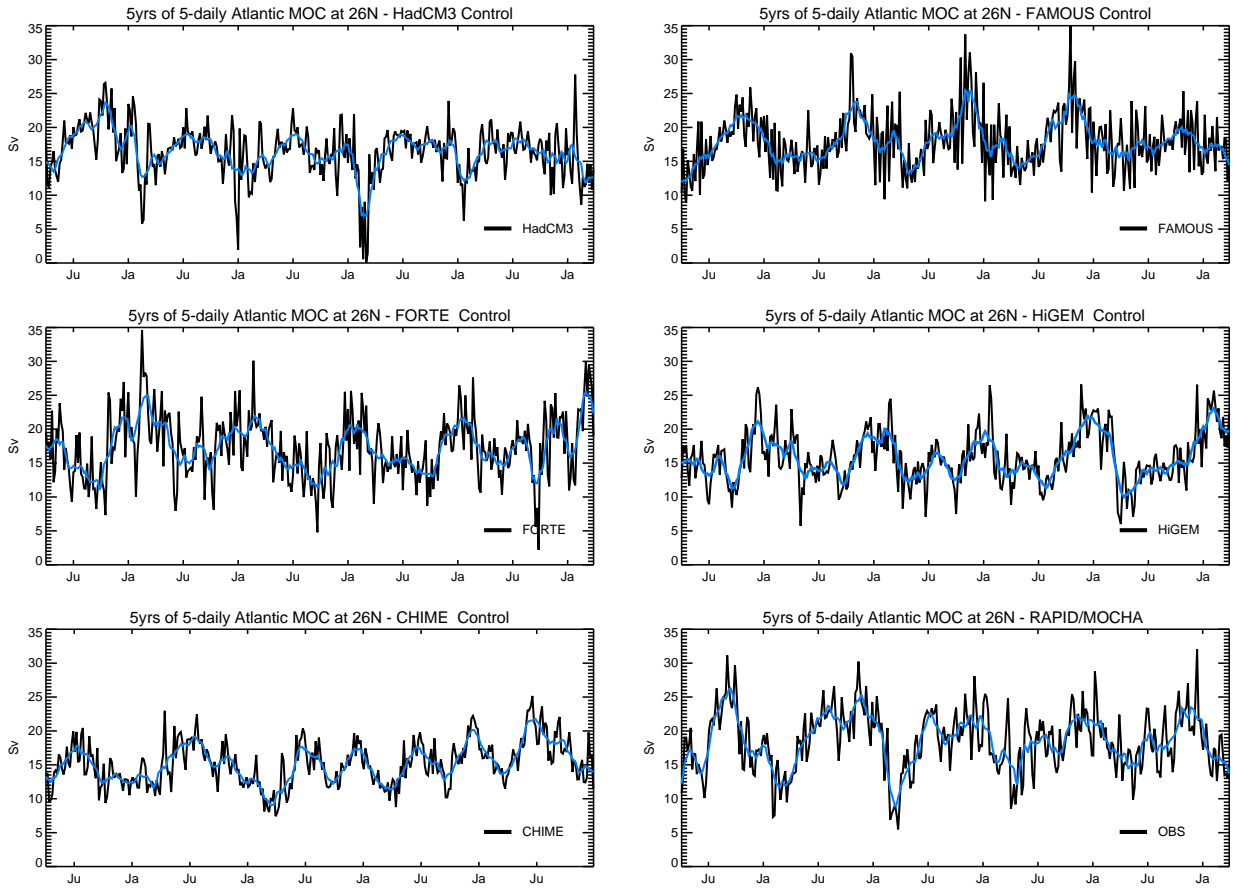


Figure 2: A 5-yr timeseries of 5-daily Atlantic MOC (T_{over}) at 26°N in observations and in the RAPID-AOGCMs. (Data with a 45-day moving average is shown in blue.) Other RAPID-models, GENIE and FRUGAL with simple atmospheric components, have little interannual variability. The last 5 years of the 10 years of data from each AOGCM is shown here.

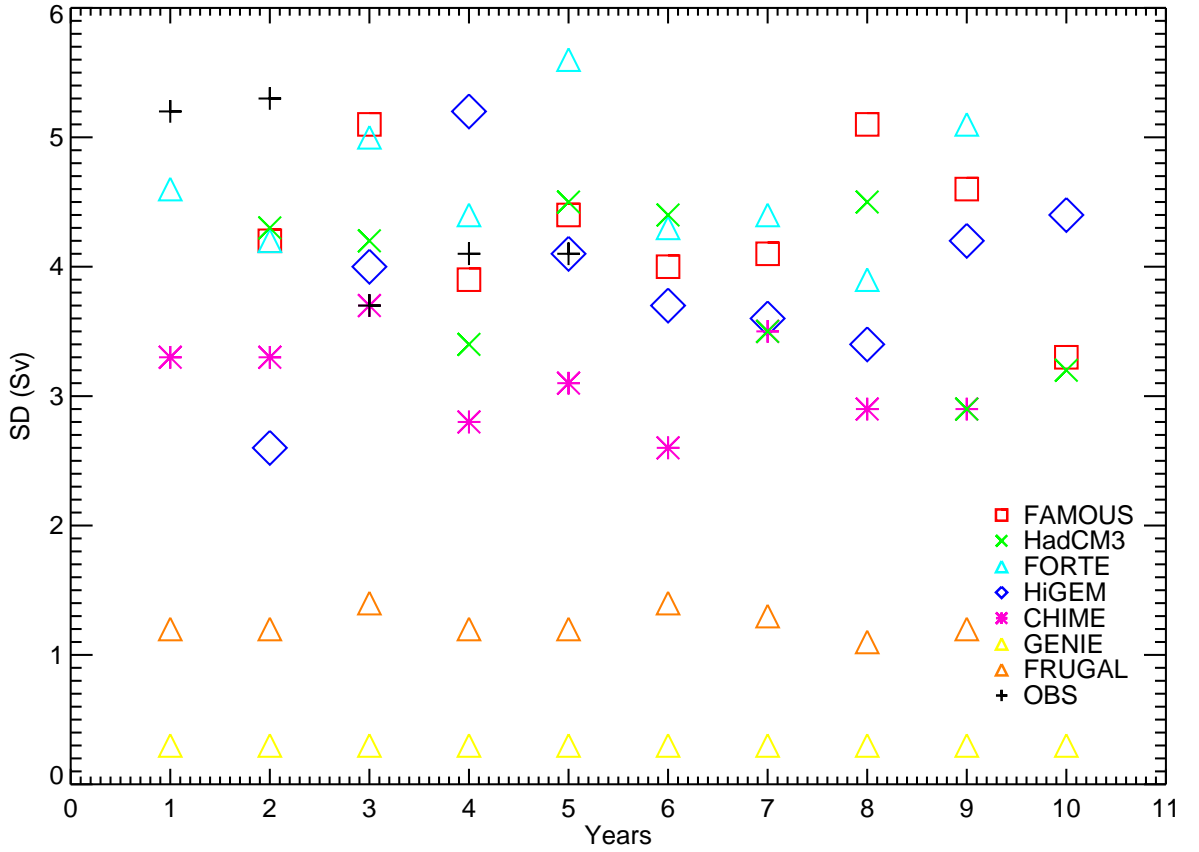


Figure 3: Standard deviation of the 5-daily Atlantic MOC (T_{over}) at 26°N for each year of simulations and observations. (Since the observational timeseries starts in April, the SD is calculated from April to March and some models are missing a year because of wanting to start all the years in April.)

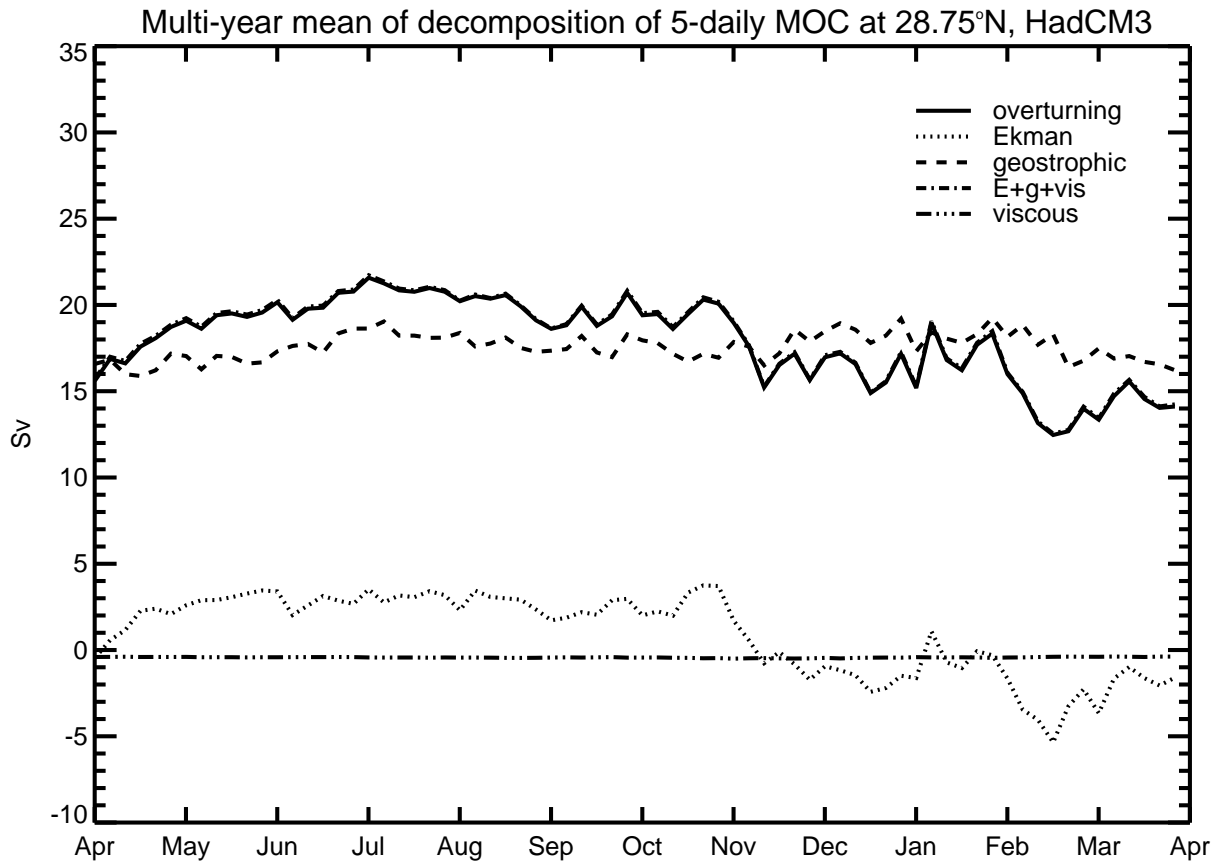


Figure 4: Decomposition of 5-daily Atlantic MOC (T_{over}) into physical components at about 29°N in HadCM3. The sum E+g+vis (dash-dotted) is almost coincident with the total overturning (solid).

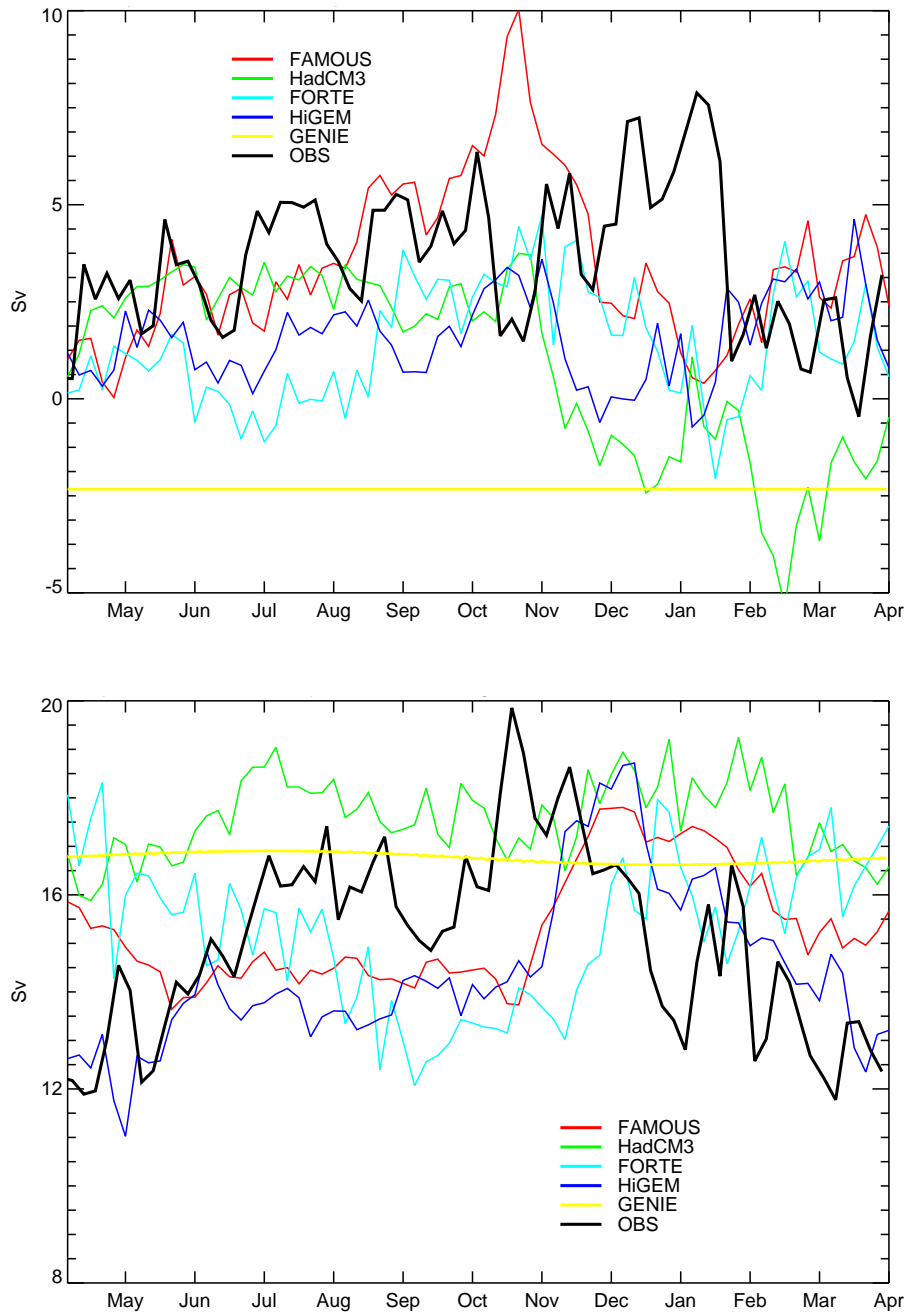


Figure 5: Annual cycle of 5-daily Atlantic MOC (T_{over}) components at about 29°N - a) Ekman component (T_{Ek}) and b) Geostrophic component (T_{geo}).

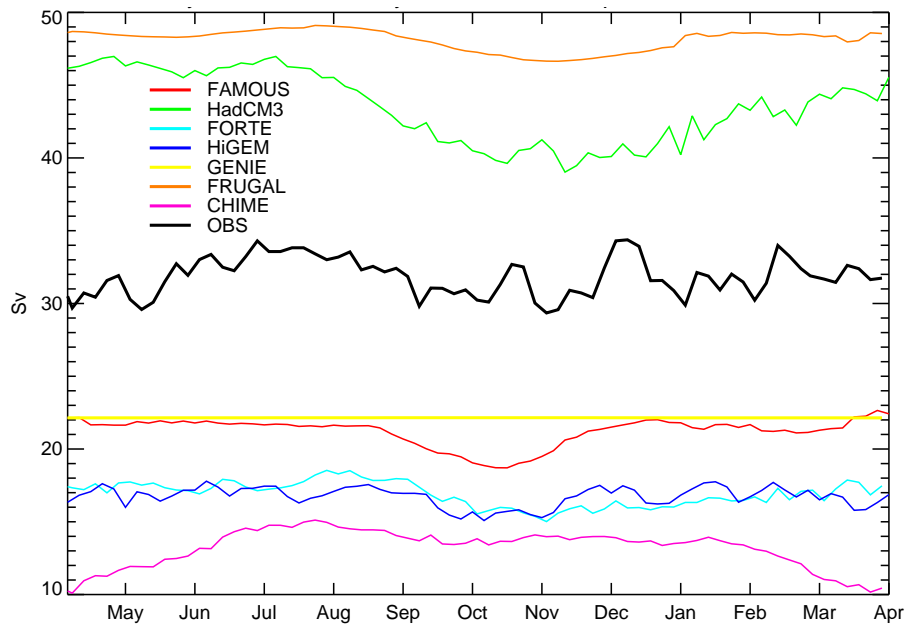


Figure 6: Annual cycle of 5-daily Western Boundary Current (T_{GS}) at about 29°N calculated geographically (See Sect. 4 for details).

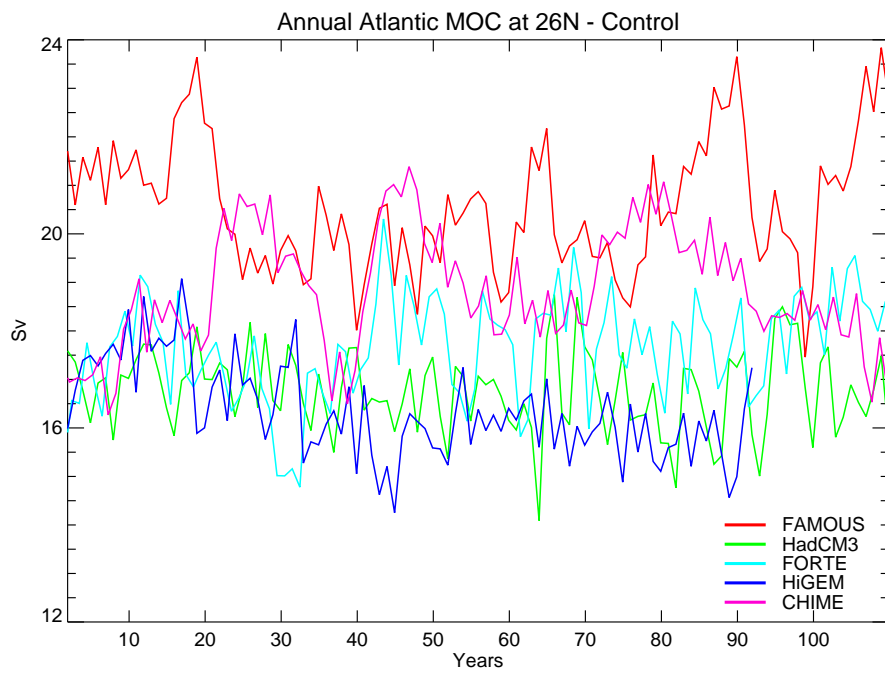


Figure 7: Annual time series of the Atlantic MOC (T_{over}) at 26°N (HiGEM data is only 90 years long after the spin-up time).

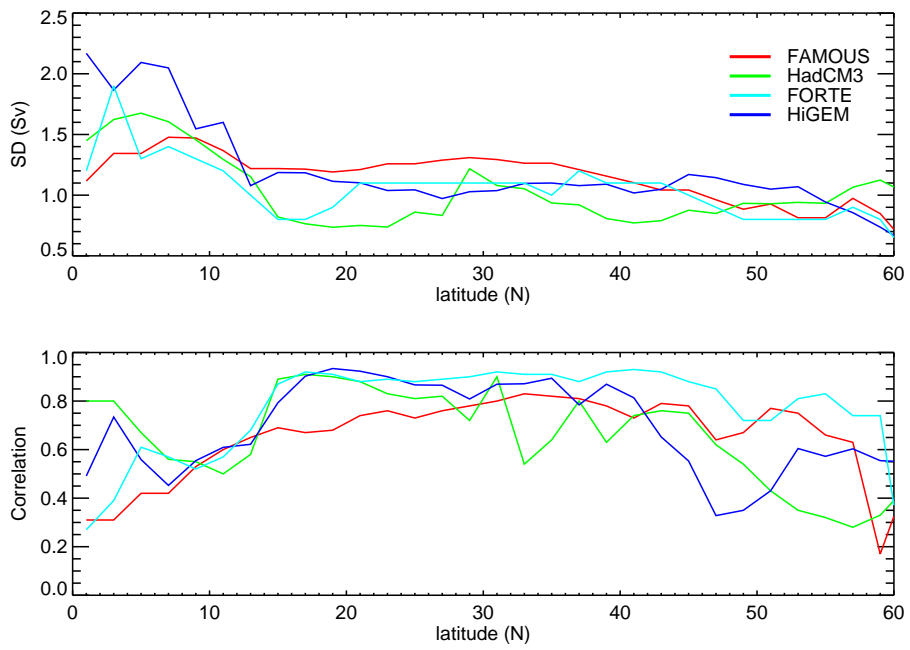


Figure 8: Zonal profile of a) annual ocean meridional overturning transport (T_{over}) variability (Sv) and b) correlation of annual T_{over} and ocean meridional heat transport in the north Atlantic.

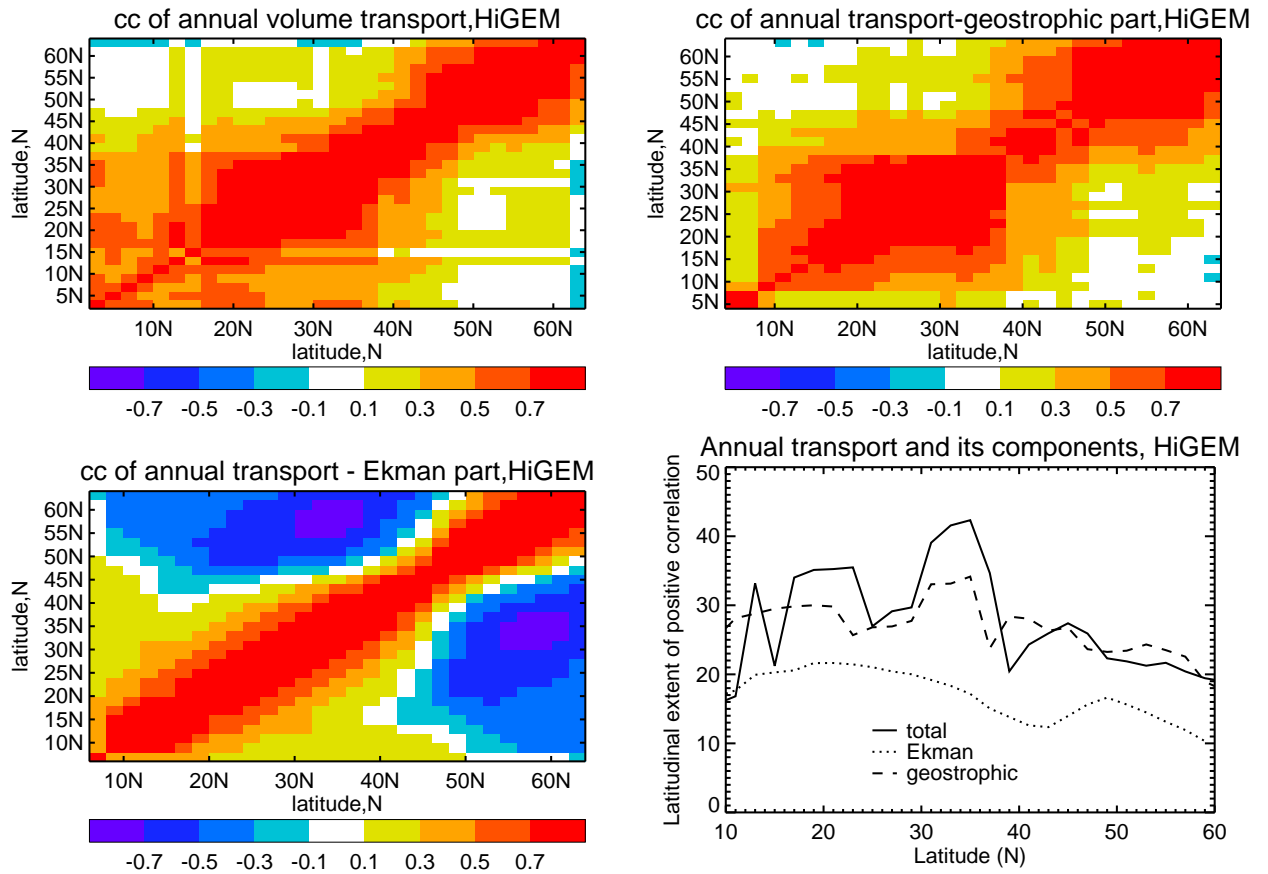


Figure 9: Cross-correlation of ocean meridional overturning transport, T_{over} and its physical components, between latitudes in the north Atlantic in HiGEM: Annual T_{over} (top left), geostrophic, T_{geo} (top right), Ekman, T_{ek} (bottom left) and their meridional correlation length (bottom right). Correlation length ($^{\circ}$ lat) as a function of latitude y is defined as the width of the range of latitudes whose timeseries which have a temporal correlation exceeding 0.5 with the timeseries at latitude y .

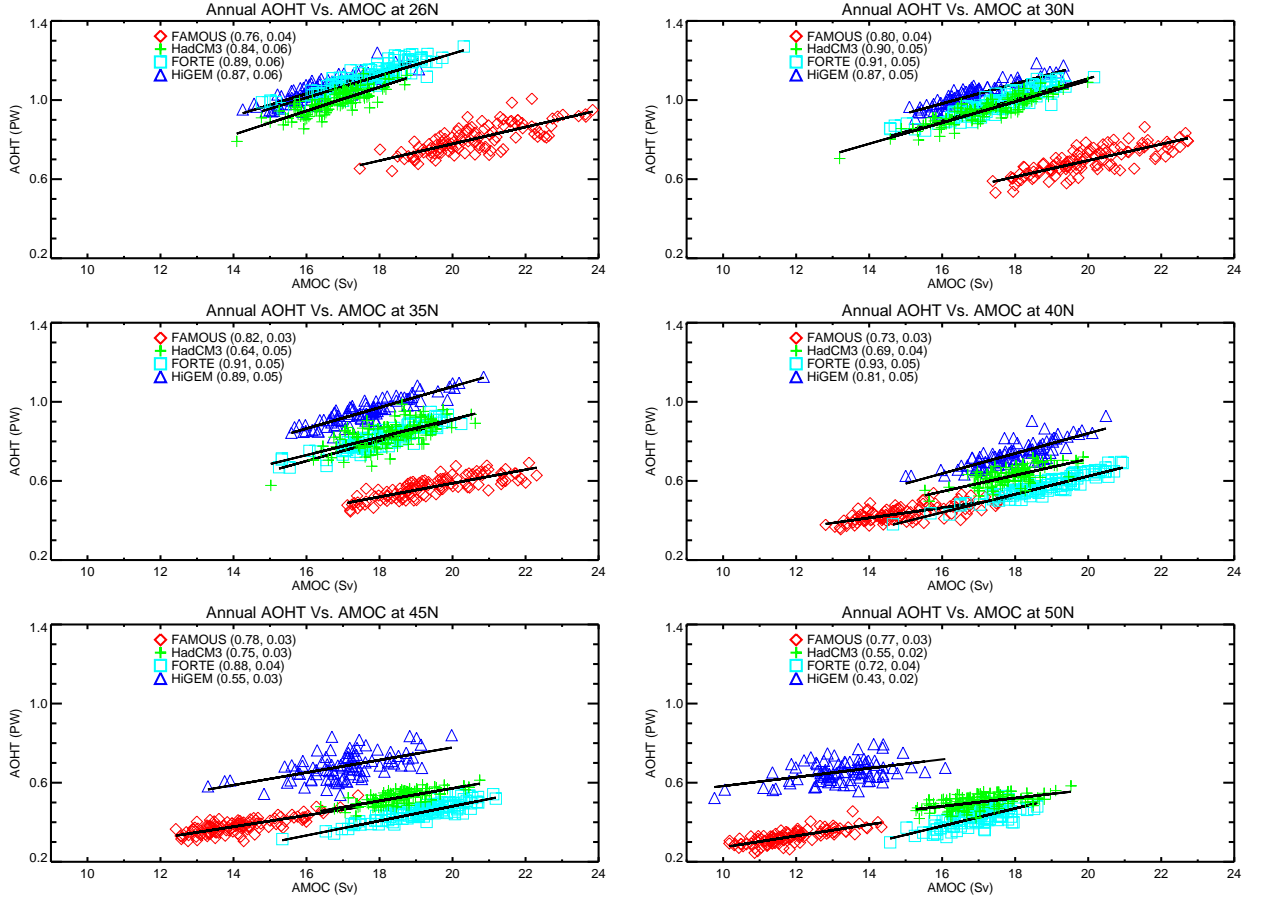


Figure 10: Scatter plot of annual-mean ocean meridional overturning transport, T_{over} (Sv) and ocean meridional heat transport (PW) at various latitudes in the north Atlantic in different models. The correlation coefficients and slopes of the regression are given in brackets.

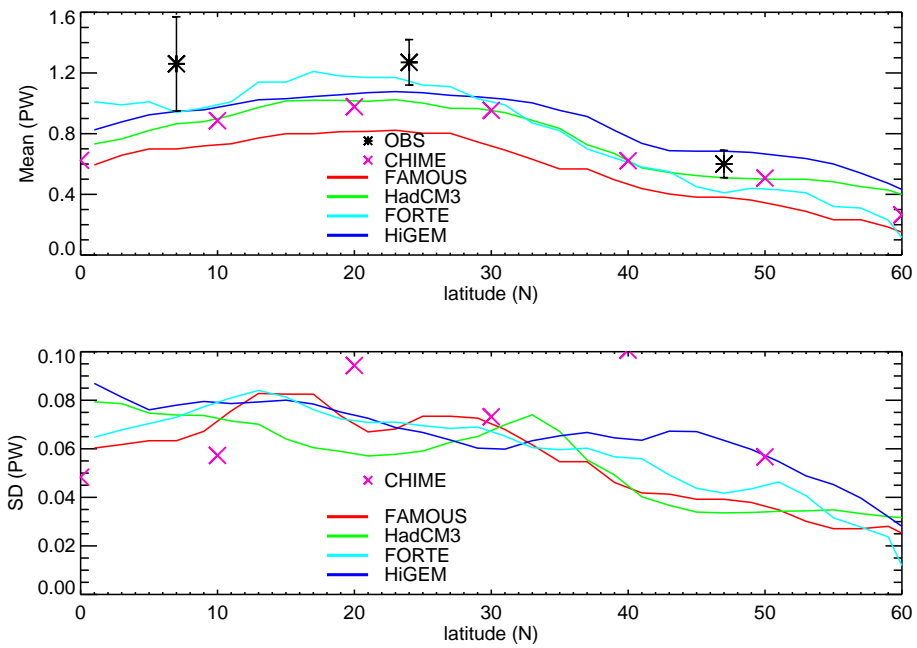


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