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North Atlantic simulations in Coordinated Ocean-ice Reference Experiments phase II (CORE-II). Part I: Mean states

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Abstract:

Simulation characteristics from eighteen global ocean–sea-ice coupled models are presented with a focus on the mean Atlantic meridional overturning circulation (AMOC) and other related fields in the North Atlantic. These experiments use inter-annually varying atmospheric forcing data sets for the 60-

year period from 1948 to 2007 and are performed as contributions to the second phase of the Coordinated Ocean-ice Reference Experiments (CORE-II). The protocol for conducting such CORE-II experiments is summarized. Despite using the same atmospheric forcing, the solutions show significant differences. As most models also differ from available observations, biases in the Labrador Sea region in upper-ocean potential temperature and salinity distributions, mixed layer depths, and sea-ice cover are identified as contributors to differences in AMOC. These differences in the solutions do not suggest an obvious grouping of the models based on their ocean model lineage, their vertical coordinate representations, or surface salinity restoring strengths. Thus, the solution differences among the models are attributed primarily to use of different subgrid scale parameterizations and parameter choices as well as to differences in vertical and horizontal grid resolutions in the ocean models. Use of a wide variety of sea-ice models with diverse snow and sea-ice albedo treatments also contributes to these differences. Based on the diagnostics considered, the majority of the models appear suitable for use in studies involving the North Atlantic, but some models require dedicated development effort.

Highlights

► Phase II of the Coordinated Ocean-ice Reference Experiments (CORE-II) is introduced. ► Solutions from CORE-II simulations from eighteen participating models are presented. ► Mean states in the North Atlantic with a focus on AMOC are examined. ► The North Atlantic solutions differ substantially among the models. ► Many factors, including parameterization choices, contribute to these differences.

Keywords: Global ocean–sea-ice modelling ; Ocean model comparisons ; Atmospheric forcing ; Experimental design ; Atlantic meridional overturning circulation ; North Atlantic simulations

1. Introduction

The Coordinated Ocean-ice Reference Experiments (COREs) were first introduced in [Griffies et al., 2009](#). The CORE framework defines protocols for performing global ocean–sea-ice coupled simulations forced with common atmospheric data sets. Therefore, the most essential element of the CORE framework is the forcing data sets developed by [Large and Yeager, 2004](#) and [Large and Yeager, 2009](#). The first phase of this project, namely CORE-I, involved using an idealized, i.e., synthetically constructed, one-year repeating cycle of forcing, referred to as normal year forcing (NYF). The primary goal was to investigate and document the climatological mean ocean and sea-ice states obtained after long (at least 500 years) integrations, with the hypothesis that global ocean–sea-ice models run under the same atmospheric state produce qualitatively similar solutions. A comprehensive analysis of the model simulations participating in CORE-I along with many other aspects of the CORE framework are presented in [Griffies et al., 2009](#), which finds that the above hypothesis is not valid in general, primarily depending on the particular diagnostic chosen.

The second phase of COREs, CORE-II, uses inter-annually varying atmospheric forcing (IAF) over the 60-year period from 1948 to 2007.¹ In the oceanographic community, the CORE-II simulations are usually referred to as *hindcast* experiments. These hindcasts provide a framework to evaluate ocean and sea-ice model

20 performance and study mechanisms of time-dependent ocean phenomena and their
21 variability from seasonal to decadal time scales for the recent past. Specifically, we
22 believe that the CORE-II hindcast experiments directly contribute to: i) evaluation,
23 understanding, and improvement of the ocean components of earth system models;
24 ii) investigation of mechanisms for seasonal, inter-annual, and decadal variability;
25 iii) attribution of ocean-climate events to forced and natural variability; iv) evalu-
26 ation of robustness of mechanisms across models; and v) bridging observations and
27 modeling, by complementing ocean reanalysis from data assimilation approaches.
28 They also provide consistent ocean and sea-ice states that can be used for initial-
29 ization of climate (e.g., decadal) prediction experiments. Some examples of recent
30 work demonstrating use and benefits of inter-annually forced simulations include
31 mechanisms and attributions studies on the mid-1990s weakening and warming of
32 the North Atlantic sub-polar gyre (SPG), e.g., Lohmann et al. (2009) and Yeager
33 et al. (2012), respectively, and studies on the link between the SPG and the Atlantic
34 Meridional Overturning Circulation (AMOC) as discussed in Hatun et al. (2005).
35 We note that, among these studies, Yeager et al. (2012) analysis utilized a CORE-II
36 hindcast simulation as well as decadal prediction experiments which were initialized
37 using ocean and sea-ice initial conditions from the CORE-II simulation.

38 In contrast to only seven participants in CORE-I, the present CORE-II effort
39 has grown considerably to eighteen participants (see Table 1 and Appendix A for a
40 list of the participating groups along with brief descriptions of models). They rep-
41 resent quite a diverse set of ocean and sea-ice models used in climate simulations.
42 For example, with respect to their vertical coordinates, although the majority of
43 the models use the traditional depth coordinate (e.g., NCAR, GFDL-MOM, NEMO-
44 based models), we have the participation of isopycnal coordinate (BERGEN and
45 GFDL-GOLD), hybrid coordinate (FSU), mass coordinate (GISS), and terrain fol-

46 lowing coordinate (INMOM) ocean models. Additionally, the solutions from the first
 47 unstructured finite element ocean model (FESOM from AWI) are included. We also
 48 welcome a data assimilation contribution (MRI-A) forced with the CORE-II IAF
 49 data sets. Inclusion of such an effort in the present study is intended to stimulate
 50 discussions and collaborations between the free-running and data assimilation ocean
 51 modeling communities as they have been working largely in isolation from each other.
 52 We believe that joint analysis of their simulations will benefit both communities by
 53 identifying robust features and physical mechanisms as well as systematic biases and
 54 shortcomings, leading to improvements in both approaches. As such, we do not use
 55 the MRI-A solutions as a benchmark to which the free-running simulations are com-
 56 pared, but rather treat it as just another contribution, commenting on its solutions
 57 when warranted. With their $\mathcal{O}(1^\circ)$ horizontal resolutions, none of the participating
 58 models can resolve eddies, i.e., they are non-eddy ocean models. We note that
 59 while some participants (e.g., NCAR, GFDL-MOM, GFDL-GOLD) represent mature
 60 efforts, some others (e.g., FSU, INMOM, AWI) are from relatively new endeavors.

61 The CORE-II simulations are being analyzed in several separate studies, each
 62 focusing on a specific aspect of the solutions, e.g., sea surface height (Griffies et al.,
 63 2013), the Southern Ocean and ventilation properties, the Arctic Ocean and sea-ice,
 64 and the South Atlantic. The current work represents one such study: an analysis
 65 of the Atlantic basin solutions with a focus on AMOC and related variables in the
 66 North Atlantic. We present our results in two companion papers. Part I (this study)
 67 documents the mean states to provide a baseline for the variability analysis presented
 68 in Part II (Danabasoglu et al., 2013).

69 Our focus on AMOC is motivated primarily by the role that it is thought to
 70 play in decadal and longer time scale climate variability, as well as in prediction of
 71 the earth’s future climate on these time scales. This is because its large heat and

72 salt transports significantly influence the climate of the North Atlantic and can even
73 impact global climate through atmospheric interactions (e.g., Sutton and Hodson,
74 2005; Hurrell et al., 2006). Essentially, an important, dynamically active component
75 of the memory of the climate system is thought to reside in AMOC. We believe
76 that the CORE-II hindcast experiments provide a framework to reconstruct AMOC
77 behavior during the recent past, complementing both observations and reanalysis
78 products. This work represents a first step towards more comprehensive studies that
79 use these hindcast simulations to study various AMOC-related questions further.

80 Our hypothesis remains similar to that of CORE-I: global ocean – sea-ice mod-
81 els integrated using the same inter-annually varying atmospheric forcing data sets
82 produce qualitatively very similar mean and variability in their simulations, but we
83 apply this hypothesis to the North Atlantic. Alternatively, we ask how similar or
84 dissimilar the solutions are from ocean – sea-ice models that are forced with the same
85 inter-annually varying atmospheric data sets and investigate reasons for differences
86 in their solutions. As we focus on the mean states in the North Atlantic in this paper,
87 one particular goal is to assess model fidelity by comparing model solutions to avail-
88 able observations, thus potentially identifying outliers. We also explore time-mean
89 relationships between AMOC and other fields such as meridional heat transports,
90 mixed layer depths, and sea-ice cover. We note that in contrast with the climato-
91 logical mean states discussed in Griffies et al. (2009) for CORE-I, our analysis is for
92 *present-day* conditions, providing the background states for the variability analysis
93 of Part II. Moreover, we have results from eighteen models – a more comprehen-
94 sive set than in Griffies et al. (2009). In addition, the present models (except FSU)
95 incorporate many improvements compared to those used in Griffies et al. (2009).
96 Therefore, differences in overall characteristics of these models between CORE-I and
97 CORE-II simulations reflect the combined effects of changes in model formulations

98 and forcing.

99 The paper is organized as follows. In section 2, we briefly describe the CORE-II
100 IAF data. The degree of equilibrium achieved by the models is assessed in section
101 3. The time-mean results for the AMOC; meridional heat transport; potential tem-
102 perature (θ), salinity (S), and density; mixed layer depth and ventilation; sea-ice;
103 and gyre transports are given in sections 4 through 9. The relationships between
104 the mean AMOC and the Labrador Sea (LS) hydrographic properties, LS sea-ice
105 extent, and Nordic Seas overflows are investigated in section 10. Section 11 includes
106 a summary and conclusions. As this paper is intended to be the primary reference
107 for the CORE-II IAF framework, brief model descriptions, CORE-II IAF experi-
108 mental protocol, and some details of the hydrological forcing and salinity restoring
109 are presented in Appendices A, B, and C, respectively. Because all models, except
110 GISS, use a distorted horizontal grid, a brief summary of how the zonal averages
111 and transports are calculated by the models is given in Appendix D. We discuss
112 an interesting sensitivity of meridional heat transport to a particular parameteriza-
113 tion (i.e., the Neptune parameterization) in the NOCS contribution in Appendix E.
114 Finally, a list of major acronyms is included in Appendix F.

115 2. CORE-II IAF Data

116 The CORE-II IAF global data sets used in this study are version 2 of the CORE
117 data sets described in Large and Yeager (2009). The input data are based on NCEP
118 reanalysis for the sea level pressure and near surface atmospheric state, i.e., vector
119 wind, temperature, specific humidity and density, and on a variety of satellite based
120 radiation, sea surface temperature (SST), sea-ice concentration, and precipitation
121 products. Some of these data are adjusted / corrected using more reliable in situ
122 and satellite measurements to address some known biases and limitations of the data.

123 Here, the data sets cover the 60-year period from 1948 to 2007. All forcing fields
124 vary for the 24-year period from 1984 to 2007. However, radiation and precipita-
125 tion before 1984 and 1979, respectively, are available only as climatological mean
126 annual cycles. The data frequencies are 6-hourly for sea level pressure, vector wind,
127 temperature, specific humidity, and density; daily for radiation; and monthly for pre-
128 cipitation. The data sets are available on a spherical grid of T62 resolution (about
129 1.9°) and they do not have leap years.

130 The river runoff data, containing river discharges at discrete river mouth locations
131 on a $1^\circ \times 1^\circ$ global grid, are also inter-annually varying at monthly frequency. They
132 are an updated version of the Dai and Trenberth (2002) and Dai et al. (2009) runoff to
133 correct for identified discrepancies and to ensure compatibility between the 12-month
134 climatological data and the inter-annual data. There are missing data for many rivers
135 since October 2004. The gaps were filled with the latest 5-year mean values, i.e.,
136 October 1999 - September 2004, for each month. The same fill procedure was used
137 to construct the entire runoff data for 2007. Finally, we added a time-invariant
138 distribution of runoff along the coast of Antarctica as continental runoff. Based on
139 the precipitation minus evaporation balance, Large and Yeager (2009) estimate this
140 runoff as 0.073 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$). This is distributed as a uniform flux along
141 the coastal points around Antarctica. It enters the ocean as a liquid, so there is no
142 prescribed calving of land ice. This new river runoff dataset has a global long-term
143 discharge of about 1.22 Sv, including Antarctica.

144 The CORE data sets are collaboratively supported by the National Center for
145 Atmospheric Research (NCAR) and the Geophysical Fluid Dynamics Laboratory
146 (GFDL) under the umbrella of the Climate Variability and Predictability (CLIVAR)
147 Working Group on Ocean Model Development (WGOMD). All data sets, codes for
148 the bulk formulae, technical report, and other support codes along with the release

149 notes are freely available at <http://data1.gfdl.noaa.gov/nomads/forms/core.html>.
150 Future releases of these data can be expected as improvements are made to the
151 data products and to our understanding of their biases and as data become available
152 for recent years (now available through 2009).

153 **3. Assessment of Equilibrium**

154 Following the CORE-II IAF experimental protocol (Appendix B; Griffies et al.,
155 2012), all the participating groups integrated their models for 300 years, correspond-
156 ing to five cycles of the forcing data. As the model solutions exhibit drift below
157 the upper ocean, this length of integration is clearly too short for investigations in-
158 volving deep ocean tracer properties that evolve on long diffusive time scales. For
159 such studies, longer integrations and / or detrending of model data may be needed.
160 In contrast, in our experience (as documented in, e.g., Doney et al., 2007; Lohmann
161 et al., 2009; Yeager et al., 2012), 300-year integration lengths are sufficient for studies
162 involving, for example, AMOC, subtropical and subpolar gyres, convection and deep
163 water formation in the North Atlantic, and upper ocean mean and variability.

164 To evaluate the degree of equilibrium achieved in the simulations, we use the
165 AMOC annual-mean maximum transport time series at 26.5°N as our metric (Fig.
166 1). This latitude is chosen as a representative latitude as we obtain qualitatively
167 similar results at several other latitudes – AMOC at 26.5°N will also be used for
168 comparisons with the RAPID observations (Rapid Climate Change mooring data,
169 Cunningham et al., 2007) later. Here, we seek to determine the repeatability of the
170 AMOC time series from one forcing cycle to the next one for each model – except
171 MRI-A because it was run for only one forcing cycle. This is quantified in Fig. 2 by
172 considering root-mean-square (rms) differences and correlations of the AMOC time
173 series of Fig. 1 for each subsequent forcing cycle pair. Specifically, for each model,

we compute rms differences and correlations between forcing cycles 2 and 1, 3 and 2, 4 and 3, and finally 5 and 4. The rms measures the differences in the means, trends, and variability from one cycle to the next one and if a model duplicates its AMOC time series identically without any trends, then the rms differences are expected to asymptote to zero. Correlations are more specific, focusing only on the repeatability of the AMOC variability during each subsequent forcing cycle pair, using detrended (and mean subtracted) time series. At equilibrium, correlations would approach unity. A major caveat in our rms and correlation analysis here is that we assume internal model variability is much smaller than the forced variability in this class of coarse resolution (viscous), non-eddyding ocean models. Otherwise, an equilibrated model would show non-zero rms and correlations of less than one. We note that our analysis excludes the first ten years of each cycle to avoid the large adjustments associated with the unphysical jump in the forcing from 2007 back to 1948.

Using an arbitrary lower limit of 0.95 for the correlation coefficients and an upper limit of 0.5 Sv for the rms differences, Fig. 2 shows that half of the participating models (NCAR, MIT, MRI-F, ACCESS, NOCS, CERFACS, CNRM, CMCC, and GFDL-GOLD) obtain a *practical* AMOC equilibrium state by the fifth forcing cycle. In some of these models, the above equilibrium criteria are satisfied even earlier by the third cycle. BERGEN and GISS also come very close to satisfying both criteria. In contrast, AWI, GFDL-MOM, ICTP, FSU, and INMOM duplicate neither the variability nor the amplitude (or mean) of AMOC transports between two consecutive cycles as also evidenced in Fig. 1. KIEL reproduces the variability between the fourth and fifth cycles, but the rms differences reflect the large upward trend seen in Fig. 1.

We will discuss the differences in AMOC transports among the models in the following sections. Here, we note that the models show a significant spread in their

200 initial AMOC magnitudes – despite very similar initialization of the ocean models
201 (see Appendix B) – and there are substantial differences in their spin-ups. Such
202 differences were also reported in Griffies et al. (2009) for the CORE-I simulations.

203 In the rest of this paper, we focus on the results from the fifth cycle of the
204 simulations. Unless otherwise noted, we define the mean states as the 20-year time-
205 means for years 1988-2007, corresponding to simulation years 281-300. We also use
206 March-mean data obtained by averaging monthly-mean March data for the same
207 20 years. For our LS analysis, we perform spatial averages in a region bounded by
208 60° - 45° W and 50° - 65° N (indicated in Fig. 8). Furthermore, in our presentation,
209 we tried to group together the results from the models with close family ties, i.e.,
210 similar ocean base codes or usage of non-level vertical coordinate systems. Thus,
211 the MOM-based models (GFDL-MOM, ACCESS, ICTP), the NEMO-based models
212 (KIEL, NOCS, CERFACS, CNRM, CMCC), and the density (BERGEN, GFDL-
213 GOLD), hybrid (FSU), mass (GISS), and sigma (INMOM) coordinate models are
214 grouped together, respectively (see Table 1).

215 In addition to AMOC spatial distributions, AMOC maximum transports at 26.5°
216 and 45° N are used as two representative latitudes, with the former latitude allowing
217 the opportunity to compare model results to those of the RAPID observations and
218 the latter latitude providing a measure of mid-latitude AMOC. We use the total
219 AMOC transports in our analysis, i.e., the sum of the Eulerian-mean, mesoscale
220 eddy, and submesoscale eddy contributions, if the latter two are available. While
221 all but one (INMOM) of the models include a variant of the Gent and McWilliams
222 (1990) parameterization to represent the advective effects of the mesoscale eddies,
223 only four models (ACCESS, GFDL-GOLD, GFDL-MOM, and NCAR) employ a
224 submesoscale eddy parameterization (Fox-Kemper et al., 2011). Because we are
225 primarily interested in large-scale sub-thermocline (below 500 m) characteristics of

AMOC and the impacts of both the mesoscale and submesoscale eddies are largely confined to the upper few hundred meters in the North Atlantic, missing subgrid-scale contributions from some models is not expected to affect our findings. For convenience, we refer to total AMOC simply as AMOC in the rest of this paper.

4. AMOC

We present the time-mean AMOC distributions in both depth and density (σ_2) space in Figs. 3 and 4, respectively (see Appendix D for a brief summary of zonal transport calculations). We note that time-mean AMOC in density space is calculated offline in most models, based on monthly-mean θ and S . Starting with the AMOC in depth space, we see that the cell associated with the North Atlantic Deep Water (NADW; clockwise circulation in the figures) shows substantial differences in its maximum transport magnitude as well as in its spatial structure among the models. Likely due to interpolation issues from sigma coordinates to depth space, the NADW cell is rather noisy in INMOM. The maximum NADW transports usually occur between 30°-45°N and broadly around 1000 m depth. There are, however, several noteworthy exceptions to these generalizations: i) the maximum transport is located further north at about 55°N in ICTP; ii) INMOM has many local maxima and small-scale circulation patterns, and iii) there are at least four local maximum transport locations in MRI-A – a feature likely resulting from internal sources and sinks of heat and salt (density) and also seen in several other ocean reanalysis products (see Munoz et al., 2011). The maximum NADW transport magnitudes are between about 8–28 Sv with FSU, NOCS, MIT, and CMCC at the low end (8–12 Sv) and NCAR and ICTP at the high end (26–28 Sv) of this range. The NADW penetration depth as measured by the depth of the zero contour line also varies significantly among models from about 2500 m in MIT and AWI to as deep as 3750–4000 m in

251 NCAR, CNRM, GISS, and MRI-A. In FSU, the NADW penetration depth is rather
 252 shallow (< 2000 m) between about 45° and 65°N . The transports associated with
 253 the Antarctic Bottom Water (AABW; counter clockwise circulation at depth in the
 254 figures) are < 6 Sv, with most models showing maximum transports of about 2–4
 255 Sv.

256 A comparison of AMOCs in depth and density space (Figs. 3 vs. 4) shows that
 257 the NADW maximum transport locations are shifted northward to about 45° - 60°N
 258 with usually similar or slightly stronger maximum transports in density space than
 259 in depth space. An exception is ICTP where the maximum transport is down from
 260 28 to 16 Sv. Another notable feature is that FSU in density space shows an even
 261 weaker maximum transport (in high density classes) than its maximum in depth
 262 space (about 4 vs. 8 Sv, respectively). Model differences displayed in Fig. 3 are also
 263 present in Fig. 4, including weaker transports for FSU, NOCS, MIT, and CMCC.

264 Figure 5 provides a quantitative comparison of the model AMOC profiles with the
 265 profile based on the RAPID data (Cunningham et al., 2007) at 26.5°N . In these plots,
 266 we use the 4-year mean for years 2004-2007 for the model data while the RAPID data
 267 represent the 4-year mean for April 2004 - March 2008. Additionally, we do not adjust
 268 the model profiles to have no net mass (or volume) transport across this latitude
 269 whereas in the RAPID analysis such a constraint was enforced. Therefore, the model
 270 profiles include relatively small ($\mathcal{O}(1$ Sv)) Bering Strait and even smaller surface
 271 freshwater flux contributions (if applicable). The profiles show the total integrated
 272 transport between the surface and a given depth, with negative and positive slopes
 273 indicating northward and southward flow, respectively. The RAPID estimate for
 274 the NADW maximum transport at this latitude is 18.6 Sv, occurring at about 1000
 275 m depth. Over this short observational record, the annual-mean AMOC maximum
 276 transports in RAPID vary by about ± 1 Sv around its mean value. This observational

277 profile, including its maximum transport, is captured remarkably well by NCAR
 278 in the upper 2000 m. The majority of the models underestimate the maximum
 279 transport with FSU showing the smallest transport with 5.5 Sv. However, several
 280 models (GFDL-MOM, KIEL, CNRM, BERGEN, GISS, and INMOM) are within
 281 10% of the RAPID maximum transport estimate. It is quite evident that the NADW
 282 penetration depth is much shallower in most of the models than in RAPID, but
 283 NCAR, MRI-A, and CNRM penetration depths come close to that of RAPID. Here,
 284 NCAR employs an overflow parameterization to represent Nordic Seas (Greenland-
 285 Iceland-Norwegian Seas) overflows (Danabasoglu et al., 2010) and MRI-A assimilates
 286 observational data. It is also clear that all models have difficulties in the AABW
 287 representation, particularly with its depth range. Associated with shallower NADW,
 288 AABW occupies a much broader depth range than in RAPID where it is confined
 289 to depths deeper than 4400 m. With the exception of NCAR, KIEL, MRI-A, and
 290 INMOM, the models have AABW maximum transports of 1–3 Sv, bracketing the
 291 RAPID estimate of about 2 Sv. In this integrated measure at this latitude, AABW
 292 maximum transport is < 1 Sv in KIEL and MRI-A; NCAR has near-zero transport;
 293 and INMOM does not show any signatures of AABW.

294 There are some similarities in the AMOC distributions between two of the MOM-
 295 based contributions (GFDL-MOM and ACCESS), but they show differences in many
 296 details. No obvious grouping of the NEMO family of models is suggested. KIEL,
 297 NOCS, CERFACS, CNRM, and CMCC show significant differences in their NADW
 298 and AABW depictions among themselves, due to differences in their parameteriza-
 299 tions, parameter choices, vertical grid levels, etc. in their ocean models and due to
 300 use of different sea-ice models.

301 Finally, we note that the present FSU contribution uses the same HYCOM (HY-
 302 brid Coordinate Ocean Model) code as in the Griffies et al. (2009) CORE-I study

303 where its AMOC transport was somewhat larger than reported here. The reasons
 304 for weaker AMOC transports with HYCOM under CORE-II forcing remain unclear.
 305 However, preliminary results from a new configuration of HYCOM show much im-
 306 proved representation of AMOC with a time-mean maximum NADW transport of
 307 >17 Sv (Rainer Bleck and Shan Sun, 2013, personal communication). This config-
 308 uration uses a different sea-ice model; employs a different reference pressure for the
 309 potential density; and advects $\theta - S$, thus preserving both heat and salt in the ocean
 310 model. We hope to include the new HYCOM version in future CORE-II studies
 311 when its integration is finalized.

312 **5. Meridional Heat Transport**

313 The Atlantic Ocean time-mean meridional heat transport (MHT) distributions
 314 from all the models are presented in Fig. 6. For comparison purposes, the figure also
 315 includes the implied transport estimates from Large and Yeager (2009) calculated
 316 using the CORE-II inter-annual fluxes and observed SSTs and sea-ice for the 1984-
 317 2006 period, and the direct estimates with their uncertainty ranges from Bryden and
 318 Imawaki (2001) and the estimate from the RAPID data (Johns et al., 2011). Within
 319 the latitude range of the maximum MHTs (10° - 30° N), the model MHTs are all lower
 320 than the mean estimates, but NCAR, AWI, GFDL-MOM, MRI-A, KIEL, CNRM,
 321 GISS, and BERGEN remain within the lower bounds of the Bryden and Imawaki
 322 (2001) estimates. They are also within or close to the lower envelope of the Large
 323 and Yeager (2009) range. None of the models is able to match the RAPID estimate
 324 range at 26.5° N. The lowest MHTs occur in MIT, MRI-F, NOCS, and CMCC, all
 325 with maximum transports of about 0.7 PW, and in FSU with a maximum transport
 326 of about 0.40 PW. (Sensitivity of MHT to the Neptune parameterization in NOCS is
 327 discussed in Appendix E.) At 11° S, while a few models (NCAR, MRI-A, and GISS)

328 produce MHTs slightly larger than the mean estimates, the other models remain
 329 below the means, but largely within the estimated uncertainty ranges. FSU is the
 330 only distribution with southward transports south of the equator in stark contrast
 331 with the other models and observationally-based data. The latitudinal variations in
 332 MHT for MRI-A reflect its AMOC structure. Such variations seem to be common in
 333 the MHT distributions obtained with some other data assimilation products as well
 334 (see Munoz et al., 2011). We believe that, as discussed in Msadek et al. (2013), errors
 335 in representations of the NADW cell and, particularly, in the vertical structure of θ
 336 (see Fig. 11), are largely responsible for the substantially lower MHTs in all model
 337 simulations compared to observational estimates even in simulations with realistic
 338 overturning strengths. Although much smaller in its contribution to MHT, errors
 339 in the gyre components can explain some of the differences as well (Msadek et al.,
 340 2013). We note that non-eddy-resolving horizontal resolutions of the present models
 341 can also contribute to low MHTs due to changes in the mean rather than the eddy
 342 heat transport (Kirtman et al., 2012).

343 At equilibrium, there is negligible storage so the positive and negative MHT
 344 slopes with respect to latitude in Fig. 6 indicate the corresponding latitude bands of
 345 zonally-integrated warming and cooling of the ocean, respectively, by the surface heat
 346 fluxes. Assuming such an equilibrium state has been achieved by the participating
 347 models, Fig. 6 implies many model differences in details of surface heat fluxes,
 348 resulting primarily from differences in simulated SSTs. One example is the much
 349 larger heat gain in BERGEN between 10° - 30° N in contrast with most of the other
 350 models where much smaller heat gains or even losses are suggested. The oceanic
 351 heat gain evident in most models between 45° - 55° N – as indicated by the positive
 352 MHT slopes – is associated with the surface heat fluxes acting to damp the cold SST
 353 biases present in these models (see Fig. 8) due to the incorrect path of the North

354 Atlantic Current (NAC) (e.g., Danabasoglu et al., 2012).

355 As hinted at above, AMOC is the dominant contributor to the Atlantic Ocean
356 MHT (Böning et al., 2001; Msadek et al., 2013). The relationship between AMOC
357 and MHT is presented in Fig. 7, considering the scatter plot of the maximum AMOC
358 transport against MHT at 26.5°N. Here and in subsequent scatter plots showing
359 AMOC strength at 26.5°N, we also include the RAPID data for reference purposes
360 only, as the model data represents the 20-year time-mean. Thus, these AMOC
361 transports do differ from those of Fig. 5. Figure 7 confirms the general tendency of
362 larger MHTs with stronger AMOC transports with a correlation coefficient of 0.89.
363 However, comparable MHTs occur for AMOC transports that differ by 2–3 Sv.
364 For example, both GFDL-MOM and AWI show similar MHTs of about 0.95 PW,
365 but their AMOC transports are about 17.8 and 14.6 Sv, respectively. We believe
366 that the larger MHT with smaller AMOC transport in AWI is primarily due to its
367 substantially larger warm biases in the upper ocean (see Fig. 11) compared to those
368 of GFDL-MOM.

369 6. Potential Temperature, Salinity, and Density

370 The time- and upper-ocean mean (0-700 m) θ , S , and in situ density model mi-
371 nus observations (World Ocean Atlas, WOA09; Locarnini et al., 2010; Antonov et al.,
372 2010) difference distributions are given in Figs. 8, 9, and 10, respectively. In many
373 regions, the θ and S differences are, to some extent, density compensating in most
374 models, as evidenced by the biases of the same signs in Figs. 8 and 9. Prominent ex-
375 amples of such biases are the warm and salty bias off the North American coast and
376 the cold and fresh bias in the mid-latitude North Atlantic present in most models.
377 These biases reach 5°-7°C and > 0.7 psu and also exist in SST and surface salinity
378 distributions (not shown). They reflect chronic model problems of the too-far-north

379 penetration of the Gulf Stream and the too-zonal NAC path compared to obser-
 380 vations. Exceptions to the cold and fresh bias associated with the too-zonal NAC
 381 path include AWI, ICTP, and INMOM where the NACs are suggested to have more
 382 northerly paths than observed. This also appears to be the case for GISS, with large
 383 positive θ and S biases in the SPG. Further north in the LS, while some models show
 384 cold and fresh biases, e.g., MIT, NOCS, and FSU, some others have warm and salty
 385 biases, e.g., NCAR, ICTP, and GISS. Similar non-uniform differences are also evi-
 386 dent in the tropical and subtropical latitudes. Most models have a salty bias near the
 387 Gibraltar Strait and off the Northwest African coast, particularly prominent in AWI,
 388 GFDL-MOM, and ACCESS. We note that ICTP shows fresh biases of > 0.7 psu in
 389 the entire Nordic Seas. We speculate that such fresh biases are likely associated with
 390 excessive sea-ice melt during the summer months, as ICTP has an extensive sea-ice
 391 cover in the Nordic Seas during the winter months (see Fig. 15).

392 The density biases, of course, reflect the θ and S biases, considering the effects of
 393 the thermal expansion and saline contraction coefficients that depend on the θ and S
 394 magnitudes (in addition to pressure). For example, at mid-latitudes, the signatures
 395 of the cold and fresh biases discussed above are present as positive density biases,
 396 indicating dominance of θ . In contrast, in the LS, the density biases appear to reflect
 397 the sign of the S biases in most models, as S changes dominate those of θ due to the
 398 smaller magnitude of the thermal expansion coefficient at low temperatures. The θ ,
 399 S , and density bias differences among the models depicted in these figures largely
 400 express the differences in the models' subtropical and subpolar gyre circulations,
 401 including differences in the Gulf Stream and NAC representations.

402 The time- and zonal-mean Atlantic Ocean θ and S model minus observations
 403 difference distributions are presented in Figs. 11 and 12, respectively (see Appendix
 404 D for a summary of zonal-mean calculations and related caveats). They also show

405 mostly same-signed θ and S differences, but there are many exceptions to this and
 406 there are many differences among the models in bias magnitudes, signs, and extents.
 407 In general, most models tend to have warm and salty biases in the upper 1000
 408 m depth and roughly south of 40°N and warm biases north of about 50°N . Several
 409 models (e.g., MIT, KIEL, CERFACS, and CNRM) show cold and fresh biases roughly
 410 between 1000–2000 m depth range and 0° – 60°N . The large fresh bias of ICTP in the
 411 upper ocean at high latitudes is clearly present in Fig. 12. Abyssal ocean biases
 412 reflect model drifts, but are usually $< 0.5^\circ\text{C}$ and 0.1 psu in magnitude. Exceptions
 413 include BERGEN and GFDL-GOLD with larger cold and fresh biases and NOCS
 414 with particularly larger warm biases. We note that GISS has larger θ biases of both
 415 signs at mid-depth and abyssal ocean, and FSU shows fresh biases at depth south
 416 of the equator. Among the models, INMOM has the most extensive and the largest
 417 magnitude warm and salty biases.

418 **7. Mixed Layer Depth and Ventilation**

419 We highlight the differences in the models' deep water formation (DWF) loca-
 420 tions by considering the March-mean mixed layer depth (MLD) distributions shown
 421 in Fig. 13 because the deepest MLDs occur in March. From among the many thresh-
 422 old criteria available to determine MLDs (see de Boyer Montégut et al., 2004), for
 423 simplicity we adopt a density-based approach where MLD is calculated as the depth
 424 at which the potential density (referenced to surface) changes by 0.125 kg m^{-3} from
 425 its surface value. We note that, for our present purposes, it is more important to use
 426 a common criterion for all models than the specific details of the MLD calculation.
 427 In those models that do not directly compute MLD online following this particular
 428 method, MLD is calculated offline using the March-mean potential density obtained

429 from the March-mean θ and S distributions. This offline method is also used to get
430 the observational MLD from the WOA09 θ and S .

431 Broadly consistent with observations, most models show essentially three DWF
432 sites identified by deep MLDs: the Nordic Seas between Iceland and Spitsbergen;
433 south of Greenland and Labrador Sea region; and south of Iceland between Greenland
434 and Scotland. Deep MLDs tend to follow the ice edge at the first two of these sites.
435 There are differences in relative depths of the deep MLD regions among the models
436 as well as between the models and those of the observations. For example, NCAR,
437 AWI, BERGEN, CERFACS, and GISS show MLDs that are deeper in the LS region
438 than in the Nordic Seas, while the opposite is evident in ACCESS, NOCS, and FSU.
439 Some of the remaining models, such as GFDL-MOM, CNRM, and GFDL-GOLD,
440 show comparably deep MLDs in their LS and Nordic Seas. The MLDs in the LS are
441 rather shallow in NOCS. In the Nordic Seas, INMOM and ICTP have the shallowest
442 MLDs. In the latter, this is due to a large fresh bias there (see Fig. 9). We note
443 that the model MLDs in LS and Nordic Seas are deeper than in observations in the
444 majority of the models.

445 To help with assessing the models' mixing processes, ventilation rates, and DWF
446 characteristics, the CORE-II protocol requests that the simulations include an ideal
447 age tracer (Appendix B). Figure 14 presents the time- and zonal-mean ideal age
448 distributions from eleven of the models that incorporated this tracer. In these distri-
449 butions, regions of low ventilation have the oldest waters while the younger waters
450 indicate recent contact with the ocean surface. We also note that, in a 300-year
451 integration, ideal age should not exceed 300 years, barring conservation issues or dis-
452 persion errors. A prominent feature in the figure is the deep penetration of young wa-
453 ters between about 50° – 70° N associated with the DWF in the North Atlantic. Using
454 the depth of the 40-year contour as a metric, the shallowest penetration depths oc-

455 cur in MRI-F, NOCS, CMCC, and INMOM with about 1000–1500 m, while NCAR,
 456 GFDL-MOM, MRI-A, GFDL-GOLD, and GISS have the deepest penetration depths
 457 of > 3500 m. These features appear to be generally consistent with the MLD dis-
 458 tributions. Another common aspect of the models is the presence of older waters
 459 – usually as a local maximum – centered at about 1000 m depth near the equa-
 460 tor. In the deep ocean, NCAR, AWI, MRI-F, NOCS, CMCC, MRI-A, BERGEN,
 461 and INMOM have ideal ages > 280 years below about 3000 – 4000 m depth, with
 462 AWI, NOCS, and CMCC showing the most extensive span of old waters. Among
 463 the models, GFDL-GOLD has the youngest deep waters with ideal ages < 240 years,
 464 indicating more vigorous mixing and ventilation of the deep oceans than in the other
 465 models. Finally, we note that significant portions of the deep ocean in INMOM show
 466 ages in excess of 300 years, suggesting either tracer conservation issues or significant
 467 dispersion errors associated with the model’s advection scheme.

468 **8. Sea-ice**

469 A detailed analysis of the North Atlantic and Arctic Ocean sea-ice solutions from
 470 these CORE-II simulations is covered in a separate study (Rüdiger Gerdes, personal
 471 communication). Here, we provide only a brief summary, focusing on the March-
 472 mean sea-ice. Because the sea-ice area (or concentration) distributions are very
 473 similar among the models for March, we show the sea-ice thickness distributions
 474 instead in Fig. 15. However, the figure can be utilized to compare the simulated
 475 sea-ice extents as approximated by the 10-cm contour line to the observational data
 476 from Cavalieri et al. (1996, updated yearly) indicated by the 15% concentration line.
 477 Overall, the majority of the models capture the observed March-mean sea-ice extent
 478 rather well. An exception is ICTP in which the Nordic Seas are largely ice covered.
 479 Although the models similarly display thicker ice in the western Arctic and increasing

480 thickness towards the Canadian Archipelago and northern Greenland, the thicknesses
481 vary considerably among the models. In about half of them (e.g., NCAR, MIT,
482 GFDL-MOM, and GFDL-GOLD), the central Arctic thicknesses are about 1.5–2 m
483 with slightly thicker ice of about 2.5–3.5 m towards the Canadian Archipelago and
484 northern Greenland. In contrast, particularly in AWI, KIEL, NOCS, CERFACS,
485 and INMOM, the thicknesses exceed 2.5 m in the central Arctic and are > 5 m
486 near the Canadian Archipelago and northern Greenland. The Arctic Ocean sea-ice
487 thickness distributions in AWI, KIEL, NOCS, and CERFACS – the latter three use
488 the same sea-ice model – are in good agreement with the very limited IceSat satellite
489 observations from Kwok et al. (2009) (not shown).

490 The sources of these model differences in sea-ice simulations are not clear and
491 a detailed analysis is beyond the scope of the present study. However, we offer
492 differences in treatments of snow on sea-ice and of subgrid-scale ice thicknesses and
493 in shortwave / albedo parameterizations as likely possibilities. Another possibility is
494 the differences in oceanic heat transport into the high latitudes and into the Arctic
495 Ocean. Our analysis, however, does not support a clear relationship between heat
496 transport magnitudes and the Arctic Ocean sea-ice area and volume, i.e., larger heat
497 transport into the Arctic Ocean does not necessarily explain reduced sea-ice (not
498 shown). We note that this finding is in contrast with a recent study by Mahlstein
499 and Knutti (2011) where a negative correlation was found between the ocean heat
500 transports at 60°N and Arctic sea-ice extents in coupled models that participated in
501 CMIP3. This discrepancy may be due to the missing feedbacks in the present ocean
502 – sea-ice simulations as detailed in Griffies et al. (2009).

503 9. Gyre Transports

504 We present the time-mean North Atlantic subtropical gyre (STG) and SPG max-
505 imum transports in Fig. 16 (left panel). These transports represent vertically-
506 integrated (barotropic) streamfunction magnitudes, thus providing measures of large-
507 scale horizontal circulations. For consistency across the models, we search for the
508 STG and SPG maximum transports between 80° - 60° W at 34° N and 65° - 40° W at
509 53° N, respectively. The SPG latitude is chosen to expedite comparisons with avail-
510 able observations (see below). For both transports, the transport values at the
511 North American coast at these latitudes are subtracted. Therefore, the maximum
512 transports are relative to the North American continent. We note that because the
513 diagnostic barotropic streamfunction fields from some models do not have constant
514 transports around continents, including North America, our diagnosed maximum
515 transports are not necessarily unique.

516 The STG transports span a range of about 17–40 Sv, with INMOM and KIEL
517 at the lower and upper ends of this range, respectively. The majority of the models
518 have STG maximum transports of 23–30 Sv. Previous studies (e.g., Bryan et al.,
519 1995) demonstrated that the dominant forcing mechanism for the STG is the wind
520 stress curl, i.e., the Sverdrup dynamics. Using the CORE-II wind stress curl with
521 the Sverdrup equation, we calculate about 23 Sv as the maximum STG transport
522 at about 34° N. The figure shows that most of the model transports are close to this
523 Sverdrup estimate. Given that the participating models are all subject to similar
524 wind stress curl forcing, we believe that the STG transport differences among the
525 models partly reflect differences in their horizontal viscosity parameterizations. We
526 note that due to the relatively coarse resolution of the models, the inertial boundary
527 currents and recirculations are largely absent in the barotropic streamfunction distri-

528 butions. Consequently, the modeled Gulf Stream and NAC transports are much less
529 than the downstream transport observations (e.g., 113 ± 8 Sv; Johns et al., 1995).

530 The SPG maximum transport range is 12–44 Sv, a broader range than in STG.
531 Here, while BERGEN and NCAR have the strongest transports, ICTP shows the
532 weakest transport. Based on observational data from Fischer et al. (2004) and Fischer
533 et al. (2010), Xu et al. (2013) report southward transport of about 37–42 Sv at the
534 Labrador Sea exit at 53°N . ACCESS, INMOM, KIEL, MRI-A, and NCAR are within
535 the estimated range. The rest of the models, except BERGEN, remain below the
536 estimates.

537 A mechanism that affects the SPG strength is the joint effect of baroclinicity
538 and relief (JEBAR; Sarkisyan and Ivanov, 1971; Holland, 1973) associated primar-
539 ily with the interaction of the dense Nordic Seas overflow waters with the sloping
540 bottom topography. Several previous studies (e.g., Böning et al., 1996; Redler and
541 Böning, 1997) implicated the characteristics of the overflow waters, e.g., density,
542 as a factor in determining the SPG strength. We show a scatter plot of the SPG
543 maximum transports against an overflow density in Fig. 16 (right panel). Here, we
544 crudely approximate this overflow density as the time-mean density of the densest
545 outflow (or southward flow) at 60°N as represented by approximately 1 Sv AMOC
546 transport in density (σ_2) space, using Fig. 4. The figure suggests no meaningful
547 connections between the overflow water densities and the SPG strengths. Although
548 a detailed exploration of the reasons for differing SPG transport magnitudes be-
549 tween the models is beyond the scope of this study, we offer differences in horizontal
550 viscosity parameterizations, sea-ice cover, and surface buoyancy fluxes as possible
551 contributors.

552 10. Relationships Between AMOC and LS Properties, Overflow Densities

553 The dense waters resulting from deep convection in the LS combine with the
554 overflow waters from the Nordic Seas (through the Denmark Strait and Faroe Bank
555 Channel) to supply the lower branch of AMOC, i.e., the NADW. In this section,
556 we briefly explore relationships between the mean AMOC transports and the LS
557 hydrographic properties, the LS sea-ice extent, and the overflow proxy density among
558 the models. We will show below that the presented relationships are consistent with
559 the following general view. The models with deeper MLDs in the LS tend to have
560 larger AMOC transports which in turn suggest higher heat and salt transports into
561 the northern North Atlantic. In such models, the LS region exhibits positive θ and
562 S biases. While the positive θ biases contribute to smaller sea-ice extents in the LS
563 region, the positive S biases tend to dominate changes in density, contributing to
564 the positive density biases in the upper-ocean, associated with the deeper MLDs.
565 However, our analysis does not distinguish, for example, if such deeper mixed layers
566 result precisely from advective fluxes (from the south) associated with AMOC itself,
567 surface buoyancy fluxes, or specifically sea-ice related changes. Thus, we do not
568 suggest a particular driving mechanism for the mean AMOC transports.

569 We first show scatter plots of the spatially-averaged θ , S , and density biases
570 against the AMOC maximum transports at 26.5° and 45°N in Fig. 17. These biases
571 are calculated in the upper 700 m for the LS region depicted in the NCAR panel
572 of Fig. 8. This region was chosen because it corresponds to a prominent DWF
573 region evident in most models (see section 7). However, we obtain very similar
574 results when we consider a broader area that includes most of the SPG region (not
575 shown). Figure 17 indicates generally larger (smaller) AMOC transports at both
576 latitudes with positive (negative) θ and S biases in the LS region. Although these

⁵⁷⁷ θ and S biases tend to partially compensate each other in their contributions to
⁵⁷⁸ density, as discussed above, density changes are largely governed by changes in S
⁵⁷⁹ as clearly evidenced in the figure. Specifically, considering the bottom panels of
⁵⁸⁰ Fig. 17, we see that MIT, ACCESS, MRI-F, NOCS, CMCC, and FSU have cold
⁵⁸¹ and fresh biases with negative density anomalies, while NCAR, ICTP, KIEL, MRI-
⁵⁸² A, BERGEN, and GISS show warm and salty biases, producing positive density
⁵⁸³ anomalies.² Thus, we find that fresh and salty LS biases are associated with weaker
⁵⁸⁴ and stronger AMOC transports, respectively. We note that while the AMOC and θ
⁵⁸⁵ bias correlation coefficients are comparable at both 26.5° and 45°N, the AMOC and
⁵⁸⁶ S bias and AMOC and density bias correlation coefficients are larger at 45°N than
⁵⁸⁷ at 26.5°N (0.74 vs. 0.60 and 0.53 vs. 0.32, respectively).

⁵⁸⁸ We next explore how the mean AMOC strength is related to the magnitude of the
⁵⁸⁹ March-mean LS MLD. Figure 18 (top panels) shows the scatter plots of the March-
⁵⁹⁰ mean LS MLDs against the mean AMOC maximum transports at 26.5° and 45°N,
⁵⁹¹ respectively. Here, the MLDs represent spatial averages calculated within the same
⁵⁹² LS region. At both latitudes, the AMOC transports vary considerably for a given
⁵⁹³ MLD, but there appears to be a tendency for larger AMOC transports with deeper
⁵⁹⁴ MLDs. Such a relationship is more prominent at 45°N than at 26.5°N as suggested
⁵⁹⁵ by the respective correlation coefficients of 0.65 and 0.52. NOCS, one of the models
⁵⁹⁶ with the weakest AMOC transports, has the shallowest average MLD in the LS or
⁵⁹⁷ south of Greenland, consistent with Fig. 13. In contrast, ICTP shows extensive and
⁵⁹⁸ deep MLDs in the LS and northern North Atlantic, with correspondingly vigorous

²In CERFACS and CNRM, the θ and S biases compensate each other and the density biases are near-zero. In contrast, the θ and S biases reinforce each other in GFDL-GOLD and INMOM. In AWI and GFDL-MOM, density biases are dictated by the S and θ biases, respectively, as the corresponding θ and S biases are near-zero.

599 AMOC at 45°N – recall that the AMOC maximum in ICTP occurs at higher latitudes
600 than in the other models. Despite an average MLD of about 500 m that is larger
601 than in MRI-F, NOCS, and INMOM, FSU has the lowest AMOC transport.

602 The scatter plots of the LS θ , S , and density biases against the LS MLDs are
603 also included in Fig. 18 (bottom panels). They show that the LS MLDs are strongly
604 dictated by the model salinity biases in the LS with a correlation coefficient of 0.87.
605 Generally, the models with salty biases tend to have deeper MLDs than the models
606 with fresh biases. The correlation coefficient between the density biases and MLD is
607 0.83 which is much larger than the correlation coefficient between the density biases
608 and the AMOC transports as the LS density changes have a more direct impact
609 on the LS MLDs. Among the models, NOCS has the shallowest MLD with a fresh
610 bias of about 0.3 psu, and ICTP has the saltiest LS with the deepest MLDs. MIT,
611 CMCC, and FSU come close to the observational MLD estimate with small density
612 biases, but such small density errors are due to the compensation of large θ and S
613 biases in density. It is interesting to note that the models appear to require positive θ
614 and S biases along with positive density and MLD biases in the LS region to achieve
615 better agreement with the observed AMOC transport at 26.5°N (e.g., NCAR).

616 In addition to the upper-ocean hydrographic properties of the LS region, the
617 Nordic Sea overflows can similarly affect AMOC as stated at the beginning of this
618 section. Indeed, several studies (e.g., Döscher and Redler, 1997; Schweckendiek and
619 Willebrand, 2005; Latif et al., 2006; Behrens et al., 2013) indicate strong connec-
620 tions between the mean AMOC maximum transports and the overflows. Specifically,
621 denser overflow waters result in higher AMOC transports, with the Denmark Strait
622 overflow as the major contributor. These findings, however, are in contrast with
623 Danabasoglu et al. (2010) and Yeager and Danabasoglu (2012) where they study im-
624 pacts of an overflow parameterization on ocean model solutions and on climate, using

625 both ocean-only simulations forced with the CORE NYF data sets and fully-coupled
 626 experiments. The parameterization produces denser overflow waters compared to
 627 control cases without this parameterization. Consequently, the NADW penetrates
 628 much deeper (as discussed in section 4), but its transport at 26.5°N changes very lit-
 629 tle and the mean AMOC maximum transport actually diminishes. Also, variability
 630 of AMOC on decadal and longer time scales is generally lower – but this reduction
 631 is not uniform in latitude and depth. These studies suggest that such reductions in
 632 the maximum transports and variability are due to the suppressed deep convection
 633 in the LS, because the denser overflow waters maintain a stratified LS.

634 The present study provides an opportunity to explore any links between the over-
 635 flow densities and the AMOC transports in the participating models. Figure 19 shows
 636 the scatter plots of the time-mean AMOC maximum transports at 26.5° and 45°N
 637 against the overflow *proxy* density described in section 9. Here, we use the AMOC
 638 transports from depth space for consistency with the previous studies. In both pan-
 639 els, the majority of the models (12) are clustered together between $36.85\text{--}37.00\text{ kg m}^{-3}$
 640 with no clear relationship between their AMOC transports and overflow densi-
 641 ties. We note that with its overflow parameterization, NCAR has one of the densest
 642 overflow waters with one of the largest AMOC transports.

643 We acknowledge that there are many caveats with this overflow vs. AMOC anal-
 644 ysis – we list a few here. First, to re-stress, our overflow density is a rather crude
 645 approximation intended to capture the overflow water densities far downstream of
 646 the sills, after most entrainment has taken place. The representation of the overflows,
 647 the bottom topography in their vicinity, and treatment of bottom flows vary quite
 648 significantly among the models. For example, NCAR uses the overflow parameteri-
 649 zation documented in Danabasoglu et al. (2010); the Denmark Strait sill depth was
 650 deepened in AWI; some models (e.g., GFDL-MOM, KIEL, MIT, NOCS) use par-

651 tial bottom cells; some models (e.g., ACCESS, CERFACS, MRI-F) employ various
652 bottom boundary layer parameterizations; or models adapt combinations of these.
653 Our results are also affected by the groups' choices of different density increments
654 when they compute AMOC in density space. Another possible explanation for the
655 lack of any clear relationship between AMOC transports and overflow densities in
656 the present set of models in contrast with some earlier studies is that these previous
657 studies were primarily concerned with sensitivities to some forcing choices in a given
658 model whereas we consider different models here.

659 We finally focus on possible links between the March-mean sea-ice cover in the
660 LS region and the previously discussed LS θ , S , and density biases as well as the
661 MLDs. These relationships are presented in Fig. 20, using scatter plots. While we
662 recognize that there are considerable spreads in all the panels, we make the following
663 general remarks. As expected, the models with colder (warmer) upper-oceans have
664 more (less) extensive ice cover in the LS with a correlation coefficient of -0.86 .
665 Models having less extensive sea-ice cover generally show salty biases. In addition to
666 advective salt fluxes associated with AMOC itself, such positive S biases may result
667 from increased evaporation due to positive θ biases in models with less ice cover,
668 exposing a broader ocean surface to colder atmospheric temperatures. We calculate
669 the observational sea-ice area for the LS region for the 1988-2007 period as 2.3×10^5
670 km^2 . Thus, the models bracket this value with eight of them below and ten of them
671 above the observational estimate. FSU emerges as an outlier with a sea-ice area
672 that is 3.5 times larger than in observations. The scatter plot of the LS sea-ice area
673 against the LS MLD (Fig. 20, bottom left) shows that as the ice cover diminishes,
674 the LS MLD tends to get deeper. Interestingly, the models with a MLD close to
675 the observationally-based estimate have much more extensive sea-ice cover than in
676 observations with the exception of INMOM. To close the loop between the variables

677 considered in this study, the final set of scatter plots (bottom middle and right)
678 show the LS sea-ice area against the AMOC transports at 26.5° and 45°N . The plots
679 confirm the general tendency of the simulations to have a stronger AMOC transport
680 with smaller LS sea-ice cover with similar (-0.77 and -0.74) correlation coefficients
681 at both latitudes. This is consistent with previous work which showed that sea-ice
682 coverage in the LS is a key factor controlling winter water mass transformation rates
683 and deep western boundary current strength (Yeager and Jochum, 2009).

684 11. Summary and Conclusions

685 We have presented an analysis of the North Atlantic Ocean solutions with a focus
686 on the mean state of the AMOC and related variables from eighteen different models
687 participating in the CORE-II effort. The associated variability study is the subject
688 of a companion paper. It is extremely pleasing to have such large and diverse world-
689 wide involvement in this endeavor, representing major modeling groups and a variety
690 of ocean and sea-ice models. In addition to the traditional level (depth) coordinate
691 ocean models, the participation of isopycnal and hybrid coordinate models, as well
692 as of models with mass (pressure) and terrain following (sigma) coordinates in the
693 vertical and of the first unstructured finite element ocean model, greatly enhanced the
694 value of this model inter-comparison effort. Furthermore, the participation of a data
695 assimilation model (i.e., MRI-A) also offers the opportunity to identify differences
696 between free-running model simulations and state estimation products.

697 As in the preceding CORE-I study (Griffies et al., 2009), we find that our starting
698 hypothesis, namely that global ocean – sea-ice models integrated using the same
699 inter-annually varying atmospheric forcing data sets will produce qualitatively similar
700 mean and variability in their simulations, is not generally satisfied for the mean states
701 in the North Atlantic. The solutions reveal significant differences among the models.

702 Not surprisingly, the model solutions also differ from available observations, but there
703 are exceptions to this generalization with some models showing good agreement with
704 observations for some diagnostics. For example, the RAPID AMOC profile, including
705 its maximum transport, is captured well in the upper 2000 m in NCAR, and some
706 other models reproduce the maximum observed AMOC transport reasonably well.
707 However, this transport is underestimated in the majority of the models. Moreover,
708 all of the models have difficulties with the representation of the AABW, and they
709 all tend to underestimate MHT.

710 The differences in the solutions do not suggest an obvious grouping of the models
711 based on their ocean model lineage. For example, the NEMO family of models
712 have significant differences in their AMOC, MLD, etc. depictions. No grouping of
713 solution properties based on model vertical coordinate representations is obvious,
714 either. Thus, we conclude that the differences in solutions among the models are
715 primarily due to the groups' use of different subgrid scale parameterizations and
716 parameter choices as well as to differences in vertical and horizontal grid resolutions
717 in the ocean models. Use of a wide variety of sea-ice models along with diverse
718 snow and sea-ice albedo treatments also contributes to differences in the solutions.
719 Such diversity in the ocean – sea-ice configurations produces differences in surface
720 buoyancy and momentum fluxes among the models particularly through differences
721 in their SSTs, despite identical atmospheric forcing data sets. We note that there are
722 undoubtedly biases in these CORE-II IAF data sets, but the present analysis does
723 not appear to expose any clear issues with forcing related to the North Atlantic.

724 Our analysis indicates that the larger AMOC transports tend to be associated
725 with deeper MLDs, resulting from increased salt content in the LS region. These
726 positive S biases occur in conjunction with reduced sea-ice cover in the LS, likely
727 due to positive θ biases. Such positive θ and S biases along with positive density

728 and MLD biases in the LS region appear to be needed by the models to match the
 729 observed AMOC transports at 26.5°N. The θ and S biases may result from advection
 730 of positive heat and salt flux anomalies (from the south) by AMOC itself, surface
 731 buoyancy fluxes, sea-ice related mechanisms, or a combination of these. In addition
 732 to the hydrographic properties and associated DWF in the LS region, the Nordic
 733 Seas overflows can also affect AMOC transports, but our study does not indicate
 734 any clear relationship between AMOC transports and an overflow proxy density. We
 735 caution, however, that the representation of overflows and the bottom topography
 736 in their vicinity vary quite significantly among the models and that our analysis is
 737 crude.

738 Regarding restoring salt fluxes, we do not find any particular links between the
 739 LS S biases and the strength (or time scale) of surface salinity restoring used by
 740 the models. For example, KIEL and BERGEN have comparable positive S biases
 741 despite their use of 1500 and 300 days, respectively, for their restoring time scales.
 742 Similarly, the negative S biases are rather similar in MIT and CMCC with restoring
 743 time scales of 1500 and 365 days, respectively. There are no apparent connections
 744 between the AMOC transport magnitudes and the surface salinity restoring strength
 745 among the models, either, even though such a relationship can exist in a given
 746 model as discussed in Appendix C, e.g., stronger restoring results in weaker AMOC
 747 transports in NCAR – in contrast with Behrens et al. (2013) where stronger restoring
 748 produces larger AMOC transports.

749 Based on the diagnostics employed here, the majority of the models appear suit-
 750 able for use in North Atlantic studies. Although all of the models will undoubtedly
 751 benefit from further improvements, a few require some dedicated development ef-
 752 fort. Considering that INMOM represents a preliminary attempt at using a sigma
 753 coordinate model in a global configuration, its solutions appear acceptable in some

754 measures, e.g., MHT, upper-ocean θ and S biases, while there are indications of larger
 755 issues in some other diagnostics, e.g., MLD, zonal-mean θ and S biases. Its subgrid
 756 scale physics can certainly be improved by including a better mesoscale eddy param-
 757 eterization, and more effort is needed to interpret its solutions and biases. Coarse
 758 model resolution, parameter choices in the ocean model, and the sea-ice model are
 759 likely responsible for the Nordic Seas fresh bias and deep MLDs in the LS in ICTP.
 760 Addressing the cold and fresh bias and associated extensive sea-ice cover problems
 761 in the LS, among others, may lead to improvements in AMOC and MHT distribu-
 762 tions in FSU. Indeed, efforts are already underway to improve HYCOM solutions
 763 by considering a new configuration of the model that advects $\theta - S$ along with a
 764 different sea-ice model and reference pressure (Rainer Bleck and Shan Sun, 2013,
 765 personal communication). Early results from this heat and salt conserving HYCOM
 766 version show much promise, including an improved representation of AMOC. Al-
 767 though providing a deeper understanding of model biases and suggesting remedies
 768 for addressing them are beyond the scope of this study, one of the basic goals of
 769 the CORE-II effort is to provide a common framework for inter-comparison of the
 770 model results and stimulate discussions and collaborations among the participating
 771 groups. We believe that such efforts are already underway as each group assesses
 772 their contributions relative to both observations and those of the other groups –
 773 as in the HYCOM example. Finally, we note that the CORE-II framework may
 774 also be adopted by the data assimilation community in their future inter-comparison
 775 projects.

776 The CORE-II experimental protocol was intended to reflect a compromise be-
 777 tween the affordability of the simulations by a broad group of researchers and the
 778 usability of the resulting solutions for scientific purposes. We believe that such a
 779 balance has been achieved as evidenced by large participation and the fidelity of the

780 simulations.

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Appendix A. Contributing Models (in alphabetical order)

Appendix A.1. ACCESS

ACCESS-OM is the ocean and sea-ice component of the Australian Community Climate and Earth System Simulator Coupled Model (ACCESS-CM; Bi et al., 2013a). ACCESS-OM comprises the NOAA/GFDL MOM4p1 ocean code (Griffies, 2009) and the Los Alamos National Laboratory (LANL) CICE4.1 sea-ice code (Hunke and Lipscomb, 2008), coupled via the CERFACS OASIS3.25 software framework (Valcke, 2006). ACCESS-OM and its performance under a CORE NYF experiment are described by Bi et al. (2013b). Details of the performance of the ocean and sea-ice components of ACCESS-OM in coupled experiments submitted to CMIP5 are given by Marsland et al. (2013) and Uotila et al. (2013), respectively.

The ocean and sea-ice components share a common horizontal orthogonal discretization having nominally 1° resolution (360 zonal by 300 meridional grid cells) with the following refinements: a tripolar grid (Murray, 1996) north of 65°N ; equatorial meridional grid refinement to $1/3^\circ$ within a band from 10°S to 10°N ; and cosine dependent (Mercator) grid cells south of 30°S to the Antarctic coast. The vertical discretization (50 layers with 20 in the top 200 m) uses the z^* geopotential coordinate (Adcroft and Campin, 2004) and partial grid cells at the bottom (Adcroft et al., 1997). Conservative temperature (McDougall, 2003) is the model’s prognostic temperature field (results presented here use diagnosed potential temperature). For the case of static instability ACCESS-OM uses explicit convection following Rahmstorf (1993). The mixed layer is represented using the K-Profile Parameterization (KPP) scheme (Large et al., 1994) with a critical Richardson number of 0.3. A

constant background vertical diffusivity ($1.0 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$) is locally enhanced by the baroclinic abyssal tidal dissipation scheme of Simmons et al. (2004), and the barotropic coastal tidal dissipation scheme of Lee et al. (2006). ACCESS-OM uses the following subgrid scale physics: isoneutral diffusion following Redi (1982); a modified Gent and McWilliams (1990) (GM) scheme following Ferrari et al. (2010) with baroclinic closure of the thickness diffusivity; and a submesoscale mixed layer restratification scheme following Fox-Kemper et al. (2011). Shelf overflows are parameterized following the sigma transport scheme of Beckmann and Döscher (1997), using the downslope mixing scheme from Griffies (2009).

The sea-ice model computes internal ice stresses by an Elastic-Viscous-Plastic (EVP) dynamics scheme (Hunke and Dukowicz, 1997), employs a layered thermodynamic scheme, uses an incremental linear remapping for estimating the ice advection, and redistributes the ice between thickness categories through ridging and rafting schemes by assuming an exponential redistribution function. Sea-ice is divided into five thickness categories with four vertical ice layers and one snow layer in each category. The ice salinity is 4 psu.

Appendix A.2. AWI

Finite Element Sea-ice Ocean Model (FESOM) is the ocean – sea-ice component of the coupled Earth System Model which is currently under development at the Alfred Wegener Institute for Polar and Marine Research (AWI). The ocean module is an unstructured-mesh model based on finite element methods and hydrostatic primitive equations (Danilov et al., 2004; Wang et al., 2008; Timmermann et al., 2009). It allows for variable mesh resolution without traditional nesting, so multiscale simulations can be conveniently conducted.

FESOM uses z -coordinates and finite element discretization with continuous lin-

ear basis functions on the A-grid. A projection method is used for solving the free surface equation, so there is no mode splitting of barotropic velocity in the model. A flux-corrected-transport advection scheme is used in tracer equations. The KPP scheme is used for vertical mixing parameterization. Both the ocean and ice modules are discretized on the same triangular surface meshes, allowing direct exchange of fluxes and fields between the two components.

The North Pole is displaced over Greenland to avoid singularity. The horizontal model resolution is nominal 1° in the bulk of the global domain, with the North Atlantic sub-polar gyre region and global coastal regions refined to 25 km. Along the equatorial band the resolution is $1/3^\circ$. In the vertical 46 levels are used, with 10 m layer thicknesses within the upper 100 m depth. The bottom topography at the Denmark Strait is deepened to 900 m, giving two cross-sill active grid points below 600 m. Biharmonic viscosity is scaled with the third power of the grid resolution, and the neutral diffusivity and GM skew diffusivity are scaled with the grid resolution. The river runoff flux is distributed around the river mouths with a linear function within 400 km distance.

Appendix A.3. BERGEN

The BERGEN contribution uses the ocean and sea-ice components of the Norwegian Earth System Model (NorESM; Bentsen et al., 2013). This model system is based on the Community Earth System Model (CESM) version 1.0.4 with the same sea-ice component and the same application of atmospheric forcing, but with a different ocean component.

The ocean component, NorESM-O, described in Bentsen et al. (2013), originates from the Miami Isopycnal Coordinate Ocean Model (MICOM; Bleck and Smith, 1990; Bleck et al., 1992), inheriting its mass conserving formulation, C-grid dis-

cretization, leap-frog time stepping for tracers and the inviscid baroclinic dynamics, forward-backward time-stepping for the barotropic equations, and momentum equations discretized in a potential vorticity/enstrophy conserving manner. The background diapycnal diffusivity is latitude dependent and increases gradually poleward from a minimum value of $10^{-7} \text{ m}^2 \text{ s}^{-1}$ at the equator. The functional latitude dependence is inspired by Gregg et al. (2003) with values of $10^{-5} \text{ m}^2 \text{ s}^{-1}$ and $1.54 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ at latitudes of 30° and 60° , respectively. Further, the background diffusivity is constrained with an upper limit of $\sim 10^{-6} \text{ m}^2 \text{ s}^{-1}$ when sea-ice is present. Shear driven diapycnal mixing follows Large et al. (1994) but with enhanced maximum diffusivity near the ocean floor to provide more realistic mixing in gravity currents. Diapycnal mixing is also driven by a fraction of the energy extracted from the mean flow by the bottom drag (Legg et al., 2006). Tidally driven diapycnal mixing follows the parameterization by Simmons et al. (2004) where the estimated conversion of tidal energy to internal waves by Jayne (2009) is used. The ocean model does not support mass exchange through the surface, thus fluxes of fresh water are converted to a virtual salt flux. The sea-ice model, in the configuration used in this study, is unaltered from the CESM version described in Appendix A.17, which is based on version 4 of the LANL sea-ice model (CICE4; Hunke and Lipscomb, 2008)

The ocean and sea-ice components share the same tripolar grid with a 1° resolution along the equator. The grid cells are optimized for isotropy except in the equatorial region where the meridional resolution approaches 0.25° . In the Southern Hemisphere the grid singularity is at the South Pole, while the two grid singularities in the Northern Hemisphere are located in Canada and Siberia. The ocean model is configured with 51 isopycnic layers referenced at 2000 db. The surface mixed layer is divided into two non-isopycnic layers.

904 *Appendix A.4. CERFACS*

905 CERFACS-ORCA1 is used as the ocean component of CNRM-CM5, the Earth
906 System Model assembled by Météo-France and CERFACS for CMIP5. It is a 1°
907 model configuration of the version 3.2 of the Nucleus for European Modelling of
908 the Ocean (NEMO) framework. As many aspects of the CERFACS setup are very
909 similar to the NOCS version detailed in Appendix A.18, we list only the differences
910 from NOCS-ORCA1.

911 There are 42 vertical levels, monotonically increasing from 10 m near the surface
912 to 300 m in the abyssal ocean. The three-waveband scheme of Lengaigne et al.
913 (2007) is run with a constant chlorophyll value of 0.005 g Chl L⁻¹. The base value
914 of vertical diffusivity is $1.2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ only poleward of 15° of latitude. Between
915 15°-5°S and 5°-15°N, it is linearly ramped down to the constant value of 1.2×10^{-6}
916 $\text{m}^2 \text{ s}^{-1}$ in the equatorial band of 5°S-5°N, following Gregg et al. (2003). A spatially
917 varying geothermal heat flux through the ocean floor with a global mean value of 86.4
918 mW s^{-2} (Emile-Geay and Madec, 2009) is applied. The discretized version of the
919 isoneutral diffusion and the GM eddy advection do not use the triad formalism. In
920 addition, a higher diffusivity of $1 \times 10^4 \text{ m}^2 \text{ s}^{-1}$ is used in the Döscher and Beckmann
921 (2000) bottom boundary layer scheme.

922 *Appendix A.5. CMCC*

923 The CMCC contribution uses the CESM framework, but the CESM ocean com-
924 ponent has been replaced with the NEMO version 3.3 (Madec, 2008). Almost all
925 aspects of the ocean model configuration are identical to those of the NOCS version
926 described in Appendix A.18. The exceptions are: i) the vertical grid has 46 levels
927 with 10 levels in the upper 100 m; and ii) the discretized version of the isoneutral
928 diffusion and the GM eddy advection do not use the triad formalism.

929 The sea-ice model CICE is the same as the one used in the CESM model (Holland
 930 et al., 2012). It runs on the same horizontal grid as the ocean while using an Arakawa
 931 B-grid. The exchange of vector fields has been carefully designed in order to properly
 932 handle the different velocity points (at the cell grid corners in the B-grid and in the
 933 middle of the cell edges in the C-grid). The other details of the ocean - sea-ice
 934 interface follow the CESM approaches except for the exchange of freshwater and salt
 935 fluxes due to sea-ice freezing and melting which follows Tartinville et al. (2001). The
 936 model time step is 1 hour and the coupling time step between the ocean and the
 937 sea-ice is 6 hours. The sea-ice model was initialized from a previous simulation. The
 938 ocean model allows water to be exchanged across the ocean surface.

939 *Appendix A.6. CNRM*

940 The major difference from the CERFACS framework described in Appendix A.4
 941 is that the sea-ice component used by CNRM-NEMO is Gelato5, not LIM2. Gelato5
 942 considers four ice thickness categories (0-0.3 m, 0.3-0.8 m, 0.8-3 m, and over 3 m).
 943 Each category has 10 vertical layers with enhanced resolution near the top of the
 944 slab. The salinity of sea-ice varies in time, based on a scheme adapted from Van-
 945 coppenolle et al. (2009). The vertical heat diffusion coefficient is a function of ice
 946 temperature and salinity, following Pringle et al. (2007). Hence, the vertical heat
 947 diffusion equation is solved by an iterative scheme. Snow aging through densification
 948 and albedo decrease are represented by a simple snow scheme (Salas-Mélia, 2002).
 949 Sea-ice dynamics is represented by the EVP scheme (Hunke and Dukowicz, 1997),
 950 and advection is based on an incremental remapping scheme described in Hunke and
 951 Lipscomb (2002). Convergence processes, which can lead to transitions between the
 952 ice categories through sea-ice rafting or ridging are represented following Thorndike
 953 et al. (1975). A more complete description of the whole ocean – sea-ice component is

954 provided by Voldoire et al. (2013). In addition, the CNRM configuration of NEMO
955 differs from the CERFACS version only by the horizontal eddy viscosity coefficient.
956 It is set to $1 \times 10^4 \text{ m}^2 \text{ s}^{-1}$ in CNRM-NEMO, consistent with the value used in the
957 CNRM-CM5.1 model (Voldoire et al., 2013).

958 *Appendix A.7. FSU*

959 The FSU contribution uses a modified Community Climate System Model version
960 3 (CCSM3) framework where the HYCOM version 2.2 (Bleck, 2002; Chassignet et al.,
961 2003; Halliwell, 2004) is employed as the ocean component in its S - density advection
962 formulation. This configuration is referred to as GLB1x3. The horizontal grid (320
963 cells in the zonal direction and 384 in the meridional direction) and topography
964 are identical to that of the CCSM3 Parallel Ocean Program (POP) except that
965 HYCOM uses staggered Arakawa-C grid while POP uses Arakawa-B grid. GLB1x3
966 is configured with 32 hybrid layers (depth or potential density) with density target
967 ranging from 28.10 to 37.25 kg m^{-3} . The model continually checks whether or not
968 grid points lie on their reference isopycnals and, if not, tries to move them vertically
969 toward the latter (Bleck, 2002). However, the grid points are not allowed to migrate
970 when this would lead to excessive crowding of coordinate surfaces. Thus, in the
971 mixed layer or in shallow water, vertical grid points are geometrically constrained
972 to remain at a fixed pressure while being allowed to join and follow their reference
973 isopycnals over the adjacent deep ocean. Therefore, HYCOM behaves like a pressure
974 coordinate model in the mixed layer or other unstratified regions, like an isopycnic
975 coordinate model in stratified regions, and like a conventional terrain-following model
976 in very shallow and / or unstratified oceanic regions (Chassignet et al., 2003, 2006).
977 The sea-ice model employed by GLB1x3 is the same version of Community Sea-Ice
978 Model (CSIM) as used in CCSM3.

979 The initial temperature and salinity are given by the Polar Science Center Hy-
 980 drographic Climatology version 3 (PHC3). The HYCOM code advects salinity and
 981 density using a second order flux corrected transport scheme. The model baroclinic
 982 and barotropic time steps are 2160 s (leap-frog) and 36 s (explicit), respectively. The
 983 model uses the KPP mixed layer sub-model (Large et al., 1994). Interface height
 984 smoothing – corresponding to Gent and McWilliams (1990) – is applied through a
 985 biharmonic operator, with a mixing coefficient determined by the grid spacing (in
 986 m) times a constant velocity scale of 0.05 m s^{-1} . For regions where the coordinate
 987 surfaces align with constant pressure (mostly in the upper ocean mixed layer), the
 988 GM parameterization is not used, and lateral diffusion is oriented along pressure
 989 surfaces rather than rotated to neutral directions. No parameterization has been
 990 implemented for overflows.

991 *Appendix A.8. GFDL-GOLD*

992 The ocean component of the GFDL-GOLD configuration employs the Generalized
 993 Ocean Layer Dynamics (GOLD) isopycnal code originally developed by Hallberg
 994 (1995) with a nominal 1° horizontal resolution refined to $1/3^\circ$ meridionally at the
 995 equator. The model includes two mixed layers, two buffer layers, and 59 interior
 996 isopycnal layers defined according to potential density referenced to 2000 dbar. The
 997 configuration is identical to that used as part of the earth system model ESM2G as
 998 detailed by Dunne et al. (2012). The GFDL-GOLD configuration uses the same sea-
 999 ice model as the GFDL-MOM configuration. Further details of how GFDL-GOLD
 1000 was configured for the CORE simulations follow that of the GFDL-MOM with two
 1001 exceptions. First, GFDL-GOLD inserts the river runoff to the nearest ocean grid
 1002 point. No further horizontal spreading is used. The model enhances energy available
 1003 for turbulent mixing at points where river water enters the ocean, so that river

1004 water is in effect mixed over the upper ocean in a manner similar to GFDL-MOM.
1005 Second, GFDL-GOLD uses a surface salinity restoring of 50 days over 50 m, which
1006 is six times stronger than the GFDL-MOM configuration. The stronger restoring in
1007 GFDL-GOLD was found necessary to retain a stable AMOC.

1008 *Appendix A.9. GFDL-MOM*

1009 The ocean component of the GFDL-MOM configuration employs the Modular
1010 Ocean Model (MOM) code from Griffies (2012) configured using a B-grid stagger-
1011 ing with the same grid resolution and bathymetry as the CM2.1 ocean component
1012 documented by Griffies et al. (2005) and Gnanadesikan et al. (2006), which was also
1013 used for the NYF simulations of Griffies et al. (2009). This grid configuration was
1014 also used in the ESM2M earth system model of Dunne et al. (2012). The grid has
1015 a nominal 1° horizontal resolution (refined meridionally to $1/3^\circ$ at the equator) and
1016 a tripolar grid poleward of 65°N . The vertical grid uses 50 levels, with 22 in the
1017 upper 220 m. The vertical coordinate is the rescaled geopotential coordinate z^* from
1018 Stacey et al. (1995) and Adcroft and Campin (2004).

1019 GFDL-MOM time steps the tracer and velocity fields using a staggered two-level
1020 scheme documented in Griffies et al. (2005) and Griffies (2004). This scheme con-
1021 serves scalar fields to within computational round-off error, with such conservation
1022 particularly important for studies of global mean sea level (see corresponding CORE-
1023 II study from Griffies et al. (2013) for discussion). Further details of the numerical
1024 methods and physical parameterizations of the ocean are provided in Griffies et al.
1025 (2005) and Dunne et al. (2012). There is one exception to the physical parame-
1026 terizations discussed in these published papers, whereby the GFDL-MOM CORE-II
1027 simulation employs a version of the Lee et al. (2006) coastal tide mixing scheme
1028 that corrects a bug, with the bug correction greatly reducing the mixing from this

1029 scheme towards more physically relevant values. Details of this bug and its correc-
1030 tion are documented in chapter 20 of Griffies (2012). The sea-ice component used
1031 in the GFDL-MOM configuration is detailed in Delworth et al. (2006), with slight
1032 modifications towards more realistic ice albedos given by Dunne et al. (2012).

1033 In these CORE-II simulations, GFDL-MOM employs a climatological chlorophyll
1034 data-set for attenuating shortwave radiation into the upper ocean. The data-set is
1035 based on an updated version of that produced in Sweeney et al. (2005), using the
1036 optical scheme from Manizza et al. (2005) for defining the shortwave attenuation.

1037 *Appendix A.10. GISS*

1038 modelER is the ocean component of the coupled NASA GISS modelE (Russell
1039 et al., 1995, 2000; Liu et al., 2003). Here, an early version of the revised E2-R code
1040 is run in stand-alone mode (Kelley et al., 2013). It employs a mass coordinate that
1041 approximates to pressure with a vertical resolution of 32 layers, ranging from about
1042 12 m at the surface to about 200 m in the abyssal ocean, and a horizontal resolution of
1043 1.25° in longitude and 1° in latitude. The model is a fully dynamic, non-Boussinesq,
1044 mass-conserving free surface ocean model. The version used here employs a linear
1045 upstream scheme for the horizontal advection of tracers and a centered difference
1046 scheme in the vertical. A 1800 s time step is used for tracer evolution.

1047 The model uses a subgrid scale parameterization to represent exchanges with
1048 unresolved straits and open ocean for up to 12 straits, e.g., the Gibraltar, Hormuz,
1049 and Nares Straits. All ocean variables are fluxed through these straits as a function of
1050 the end-to-end pressure gradients, balanced against a drag proportional to the width
1051 of the straits. The latter serves as a tuning parameter to get reasonable fluxes.

1052 modelER uses the GISS vertical mixing scheme (Canuto et al., 2010) which mod-
1053 els diapycnal mixing throughout the whole depth of the ocean, including turbulence

1054 generated by convection and shear in the mixed layer, double-diffusive effects, mixing
 1055 due to internal waves in the interior of the ocean, and mixing due to tidal interactions
 1056 with topography near the ocean bottom. Mesoscale eddies are represented by the
 1057 GM scheme coded with the skew flux formulation (Griffies, 1998) with a new three-
 1058 dimensionally varying surface-enhanced mesoscale diffusivity based on a theoretical
 1059 prediction of the surface eddy kinetic energy (Canuto et al., 2013).

1060 Sea-ice dynamics, thermodynamics, and ocean – sea-ice coupling are represented
 1061 as in the CMIP5 modelE configuration (Schmidt et al., 2013), albeit with ice on the
 1062 ocean model grid rather than that of the atmosphere. Surface turbulent fluxes over
 1063 sea-ice are calculated using the CORE prescription of transfer coefficients.

1064 *Appendix A.11. ICTP*

1065 The ICTP-MOM ocean – sea-ice model is a coarse resolution version of the GFDL-
 1066 MOM model. The model uses the z^* -coordinate ocean code MOM4p1 documented
 1067 by Griffies (2009) and the GFDL Sea Ice Simulator (SIS) sea-ice model (see more
 1068 details in Delworth et al., 2006). The model grid uses 180 cells in the zonal direc-
 1069 tion (2°), 96 latitudinal cells (1° at the equator), and 30 vertical levels with partial
 1070 step bottom topography. The model updates the tracer and baroclinic velocity with
 1071 a 9600 s time step for both inviscid dynamics and dissipative physics. Mesoscale
 1072 eddy-induced transports are parameterized following the boundary-value problem
 1073 approach of Ferrari et al. (2010), in which the variable eddy-induced advection coef-
 1074 ficient is bounded between 600 and $1400 \text{ m}^2 \text{ s}^{-1}$. Neutral diffusivity (Redi, 1982) has
 1075 a value of $800 \text{ m}^2 \text{ s}^{-1}$. The ocean model uses background vertical diffusivity values
 1076 following Bryan and Lewis (1979), with values of 0.3×10^{-4} and $1.4 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in
 1077 the upper and deep ocean, respectively. Submesoscale and overflow mixing schemes
 1078 are not implemented in this model.

1079 *Appendix A.12. INMOM*

1080 The Institute of Numerical Mathematics (INM) Ocean Model (INMOM) is the
1081 ocean component of the INM Earth Climate Model (INMCM4.0; Volodin et al.,
1082 2010). INMOM is a sigma-coordinate ocean model. It uses a displaced North Pole
1083 where the grid pole is placed in Taimyr Peninsula. There are 360 zonal and 340
1084 meridional grid cells, corresponding to 1° and 0.5° resolution, respectively. In the
1085 vertical, it employs 40 non-uniform sigma levels. The tracer equations use isopycnal
1086 diffusion with a constant mixing coefficient of $100 \text{ m}^2 \text{ s}^{-1}$, but no additional param-
1087 eterization for mesoscale eddies is used. Vertical mixing is parameterized with the
1088 Pacanowski and Philander (1981) scheme. The sea-ice model is described in Yakovlev
1089 (2009) and contains many aspects of Hunke and Dukowicz (1997) and Briegleb et al.
1090 (2004).

1091 *Appendix A.13. KIEL*

1092 The Kiel ocean model configuration ORCA05 is based on the NEMO code (version
1093 3.1.1; Madec, 2008) and belongs to the DRAKKAR framework (The DRAKKAR
1094 Group, 2007). It uses a global ocean setup coupled with a Hibler-type sea-ice model
1095 (LIM2; Fichefet and Maqueda, 1997) in a tripolar grid configuration with a nominal
1096 0.5° horizontal resolution and 46 levels in the vertical (Biaostoch et al., 2008). The
1097 layer thicknesses vary from 6 m at the surface to about 250 m in the deep ocean. For
1098 the bottom cell, a partial cell approach is used which, in combination with advanced
1099 advection schemes, leads to an improved circulation (Barnier et al., 2006).

1100 The turbulent vertical mixing is simulated with a 1.5-level turbulent kinetic en-
1101 ergy scheme (Blanke and Delecluse, 1993). Momentum equations use a bi-Laplacian
1102 horizontal viscosity. The parameterizations of isoneutral diffusion and the GM eddy
1103 advection for tracers use the same formulation and parameters as in NOCS described

1104 in Appendix A.18. For tracer advection, a total variance dissipation scheme (Zale-
1105 sak, 1979) is employed.

1106 *Appendix A.14. MIT*

1107 The MIT simulation uses the Massachusetts Institute of Technology general cir-
1108 culation model (MITgcm; Marshall et al., 1997; Adcroft et al., 2004). Aside from
1109 the CORE-II forcing and mixing parameters used here, the model setup is from the
1110 latest Estimating the Circulation and Climate of the Ocean (ECCO) framework and
1111 it is used to improve upon the estimates of Forget (2010) and Wunsch and Heimbach
1112 (2007). However, none of the ECCO optimized forcing and mixing is used in the
1113 present simulations.

1114 In the vertical, the grid consists of 50 depth levels, with 10 m grid spacing near
1115 the ocean surface, and partial step bottom topography. In the horizontal, the so-
1116 called latitude-longitude-cap grid is used. Nominal grid spacing is 1° . While the
1117 grid follows longitude and latitude lines at mid-latitudes, it turns into a quadripolar
1118 mesh over the Arctic, where the 4 model grid poles are conveniently placed on land.
1119 Vertical mixing is parameterized by a background diffusivity of $10^{-5} \text{ m}^2 \text{ s}^{-1}$, a basic
1120 convective mixing scheme, and the schemes of Gaspar et al. (1990) and Duffy et al.
1121 (1999) under sea-ice. Tracers are further mixed along isopycnals (Redi, 1982), and
1122 advection by eddies is parameterized according to Gent and McWilliams (1990). The
1123 corresponding isopycnal and thickness diffusivities are both $500 \text{ m}^2 \text{ s}^{-1}$. The sea-ice
1124 model is a dynamic / thermodynamic model with a viscous-plastic (VP) rheology
1125 following Hibler (1979). The CORE-II surface hydrological forcing is applied as water
1126 fluxes, as opposed to virtual salt fluxes.

1127 *Appendix A.15. MRI-A (assimilation, MOVE/MRI.COM)*

1128 MOVE/MRI.COM CORE-II version is a global ocean data assimilation system
1129 based on the Multivariate Ocean Variational Estimation / Meteorological Research
1130 Institute Community Ocean Model (MOVE/MRI.COM; Usui et al., 2006; Fujii et al.,
1131 2012). This system uses the same MRI.COM version with identical grid resolution,
1132 physical schemes, and parameter settings as in MRI-F described in Appendix A.16.

1133 MOVE/MRI.COM adopts a 3-dimensional variational (3DVAR) analysis scheme
1134 based on Fujii and Kamachi (2003), in which coupled temperature - salinity (θ and
1135 S) empirical orthogonal function modal decomposition is applied to the background
1136 error covariance matrix. In the system, suboptimal θ and S analysis fields above 1750
1137 m depth for a target month are estimated from the model forecast and observational
1138 data through the 3DVAR scheme, and reflected on the model fields by incremental
1139 analysis updates (Bloom et al., 1996). The system is further improved by adopting
1140 a variational quality control scheme (Fujii et al., 2005), a sequential bias correction
1141 scheme (Fujii et al., 2009), and a first-guess-at-appropriate-time scheme (Lorenc and
1142 Rawlins, 2005).

1143 In the reanalysis run, only in-situ θ and S observational profiles (including data
1144 from mooring buoys and profiling floats) are assimilated into the model. No satellite
1145 data are used to avoid data gaps. The θ and S profiles are obtained from the World
1146 Ocean Data 2009 (Boyer et al., 2009) and the Global Temperature and Salinity Profile
1147 Program (GTSP) database (Hamilton, 1994). The system also blends a monthly
1148 θ and S climatology based on the WOA09 (Locarnini et al., 2010; Antonov et al.,
1149 2010) into the model forecast before it is used in the 3DVAR scheme to suppress
1150 the deviation of the model fields from the climatology. This procedure is roughly
1151 equivalent to relaxation with a restoring time of 100 months.

1152 The MOVE/MRI.COM is run only for 70 years, starting from model year 231 of

the MRI-F integration. The first ten years of this integration is treated as a spin-up phase during which a stronger blending of observed climatology into the model forecast (equivalent to a relaxation time scale of 20 months) than the one applied during the actual integration is used to reduce biases prior to the start of the latter. Thus, the actual MRI-A integration, assimilating data during the 1948-2007 period, begins at model year 241 and essentially corresponds to the fifth forcing cycle.

Appendix A.16. MRI-F (free running, MRI.COM)

MRI.COM is the ocean – sea-ice component of MRI-CGCM3 (MRI Coupled General Circulation Model version 3; Yukimoto et al., 2011, 2012) and is based on the MRI.COM version 3 (Tsujino et al., 2010, 2011). MRI.COM3 is a free-surface, depth-coordinate ocean – sea-ice model that solves the primitive equations using Boussinesq and hydrostatic approximations. A split-explicit algorithm is used for the barotropic and baroclinic parts of the equations (Killworth et al., 1991). Horizontal resolutions are 1° in longitude and 0.5° in latitude. The horizontal grid is tripolar as prescribed by Murray (1996). The model ocean consists of 50 vertical levels with 30 in the upper 1000 m. The vertical levels shallower than 32 m follow the surface topography as in sigma-coordinate models (Hasumi, 2006). There is a bottom boundary layer (BBL; Nakano and Sugimotohara, 2002) with a 50 m thickness. The BBL is only added in the northern North Atlantic (between 50° - 70° N and 60° W- 0°) and the Southern Ocean around Antarctica (south of 60° S).

The generalized Arakawa scheme as described by Ishizaki and Motoi (1999) is used to calculate the momentum advection terms. The tracer advection scheme is based on conservation of second order moments (Prather, 1986). Mixing along neutral surfaces caused by eddy stirring is parameterized using an iso-neutral mixing coefficient of $1000 \text{ m}^2 \text{ s}^{-1}$ (Redi, 1982) and the Gent and McWilliams (1990) parameterization with

1178 a mixing coefficient of $300 \text{ m}^2 \text{ s}^{-1} \times \sqrt{\text{grid area}}/100 \text{ km}$ where grid area is in km^2 . The
1179 maximum allowed slope of iso-neutral surfaces is set to $1/1000$. The Smagorinsky
1180 (1963) horizontal viscosity formulation is applied using a flow-dependent anisotropic
1181 tensor (Smith and McWilliams, 2003) to reduce the viscosity in the direction normal
1182 to the flow. Vertical mixing is based on a generic length scale model with parameters
1183 recommended by Umlauf and Burchard (2003) with a background three-dimensional
1184 distribution following Decloedt and Luther (2010).

1185 The sea-ice component is based on Mellor and Kantha (1989). For categorization
1186 by thickness, ridging, rheology, and albedo, those of the LANL sea-ice model (CICE;
1187 Hunke and Lipscomb (2008)) are adopted with some modifications for albedo. Short-
1188 wave radiation is partitioned with a fixed ratio: 0.575 for visible and 0.425 for near
1189 infrared. The dry and wet albedos for ice are 0.8 and 0.58, respectively. Fractional
1190 area, snow volume, ice volume, ice energy, and ice surface temperature of each thick-
1191 ness category are transported using the multidimensional positive definite advection
1192 transport algorithm (MPDATA) of Smolarkiewicz (1984).

1193 *Appendix A.17. NCAR*

1194 The NCAR contribution uses the Parallel Ocean Program version 2 (POP2; Smith
1195 et al., 2010) and the sea-ice model version 4 (CICE4; Hunke and Lipscomb, 2008).
1196 They are, respectively, the ocean and sea-ice components of the Community Climate
1197 System Model version 4 and Community Earth System Model version 1 (CCSM4
1198 and CESM1, respectively; Gent et al., 2011). Here we give brief summaries and refer
1199 to Danabasoglu et al. (2012) and Holland et al. (2012) for further details.

1200 POP2 is a level-coordinate model, using the hydrostatic and Boussinesq approx-
1201 imations. A linearized, implicit free-surface formulation is employed. The global
1202 integral of the ocean volume remains constant because the freshwater fluxes are

1203 treated as virtual salt fluxes. The model uses a displaced North Pole grid with a
1204 nominal 1° horizontal resolution. The meridional resolution is increased to 0.27° near
1205 the equator. There are 60 vertical levels, monotonically increasing from 10 m in the
1206 upper ocean to 250 m in the deep ocean.

1207 A new overflow parameterization of density driven flows (Danabasoglu et al.,
1208 2010; Briegleb et al., 2010) is used to represent the Denmark Strait, Faroe Bank
1209 Channel, Ross Sea, and Weddell Sea overflows. The model tracer equations use the
1210 GM isopycnal transport parameterization in its skew-flux form (Griffies, 1998). The
1211 effects of diabatic mesoscale fluxes within the surface diabatic layer are included
1212 via a simplified version of the near-boundary eddy flux parameterization of Ferrari
1213 et al. (2008), as implemented by Danabasoglu et al. (2008). Both the thickness and
1214 isopycnal diffusivity coefficients vary identically in the vertical, following Ferreira
1215 et al. (2005) and Danabasoglu and Marshall (2007). In the upper ocean, enhanced
1216 diffusivity values are used which can be as large as $3000 \text{ m}^2 \text{ s}^{-1}$. They diminish to
1217 $300 \text{ m}^2 \text{ s}^{-1}$ by a depth of about 2000 m. In the surface diabatic layer, the horizontal
1218 diffusivity coefficient is also set to $3000 \text{ m}^2 \text{ s}^{-1}$. The restratification effects of finite-
1219 amplitude, submesoscale mixed layer eddies are included, using the mixed layer eddy
1220 parameterization of Fox-Kemper et al. (2008) and Fox-Kemper et al. (2011). The
1221 momentum equations use the anisotropic horizontal viscosity formulation in its gen-
1222 eralized form (Smith and McWilliams, 2003; Large et al., 2001; Jochum et al., 2008).
1223 The vertical mixing is parameterized using the KPP scheme (Large et al., 1994) as
1224 modified by Danabasoglu et al. (2006) with a latitudinally varying background dif-
1225 fusivity. The abyssal tidal mixing parameterization of St. Laurent et al. (2002) and
1226 Jayne (2009) is used to represent the deep vertical mixing arising from the breaking
1227 of tidally-generated internal waves over rough topography.

1228 CICE4 shares the same horizontal grid as POP2. It includes EVP dynamics

1229 (Hunke and Dukowicz, 2002), energy-conserving thermodynamics (Bitz and Lip-
 1230 scomb, 1999), and a subgrid-scale ice thickness distribution (ITD; Thorndike et al.,
 1231 1975). A fundamental improvement in the sea-ice component is the incorporation of
 1232 a new radiative transfer scheme for the treatment of solar radiation (Briegleb and
 1233 Light, 2007; Holland et al., 2012). This scheme calculates multiple scattering of solar
 1234 radiation in sea-ice using a delta-Eddington approximation with inherent (i.e., mi-
 1235 croscopic) optical properties that specify scattering - absorption properties for snow,
 1236 sea-ice, ponds, and included absorbers. The resulting surface albedo and absorbed
 1237 shortwave flux are computed using this new radiative transfer scheme. Hence the
 1238 surface albedos are not directly tuned and instead the inherent optical properties
 1239 of snow, bare sea-ice, and melt ponds are adjusted within two standard deviations
 1240 of the observations taken during the Surface Heat Budget of the Arctic (SHEBA)
 1241 experiment in 1997-1998.

1242 *Appendix A.18. NOCS*

1243 We note that an expanded description of the NEMO framework is only provided
 1244 here to serve as a reference for other models using the same framework.

1245 NOCS-ORCA1 is the 1° model configuration of the NEMO 3.4 framework being
 1246 used at the National Oceanography Centre Southampton (NOCS). It is a z -level
 1247 Boussinesq global coupled ocean – sea-ice model. NOCS-ORCA1 includes the ocean
 1248 circulation model OPA (Madec, 2008) coupled to the Louvain-la-Neuve Ice Model
 1249 sea-ice model LIM2 (Timmermann et al., 2005), but with EVP instead of VP ice
 1250 rheology (Hunke and Dukowicz, 1997) on the C-grid (Bouillon et al., 2009). The
 1251 horizontal mesh is tripolar (Timmermann et al., 2005; Hewitt et al., 2011), based
 1252 on a 1° Mercator grid, but with additional refinement of the meridional grid to $1/3^\circ$
 1253 near the equator. North of 20°N the grid starts to deviate from Mercator as a result

1254 of the tripolar grid, but does not differ significantly until 60°N. Over the Arctic
 1255 Ocean, the model resolution is about 50 km. Model level thicknesses are about
 1256 1 m near the surface, increasing to about 200 m at 6000 m depth with 19 levels
 1257 in the upper 50 m and 25 levels in the upper 100 m. Topography is represented
 1258 with partial cells (Barnier et al., 2006). A linear free-surface formulation is employed
 1259 (Roullet and Madec, 2000), where lateral fluxes of volume, tracers and momentum are
 1260 calculated using fixed reference ocean surface height. Temperature and salinity are
 1261 advected with the total variance dissipation scheme (Cravatte et al., 2007), a second-
 1262 order, two-step monotonic scheme with moderate numerical diffusion. An energy and
 1263 enstrophy conserving scheme (Le Sommer et al., 2009) is used for momentum.

1264 Precipitation and evaporation are effected by volume input through the ocean
 1265 surface; therefore, they affect the sea surface height as a volume flux and the salinity
 1266 as a concentration / dilution term. Salinity is also restored by volume input. The
 1267 global mean of freshwater budget is set to zero at each model time step. Ice melting
 1268 and freezing instead drive salt fluxes through the ocean surface calculated assuming
 1269 constant ice (6 psu) and ocean (34.7 psu) salinities in order to conserve salt during
 1270 the ice freezing / melting cycle.

1271 Shortwave radiation is attenuated using the chlorophyll-dependent three-waveband
 1272 (RGB) scheme of Lengaigne et al. (2007) together with an observed (seasonally and
 1273 spatially varying) chlorophyll climatology (SeaWiFS, averaged 1999-2005). Momen-
 1274 tum and tracers are mixed vertically using a turbulent kinetic energy (TKE) scheme
 1275 (Madec, 2008) based on the model of Gaspar et al. (1990). It also includes a Lang-
 1276 muir cell parameterization (Axell, 2002), a surface wave breaking parameterization
 1277 (Mellor and Blumberg, 2004), and uses an energetically consistent time and space
 1278 discretization (Burchard, 2002; Marsaleix et al., 2008). Base values of vertical diffu-
 1279 sivity and viscosity are 1.2×10^{-5} and $1.2 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, respectively. Tidal mixing is

parameterized following Simmons et al. (2004), using an internal wave energy field derived from the output of the barotropic global ocean tide model MOG2D-G (Carrère and Lyard, 2003). In addition, the Koch-Larrouy et al. (2007) parameterization for tidal mixing is used in the Indonesian area.

Lateral diffusivity is parameterized by an iso-neutral Laplacian operator with a coefficient of $1000 \text{ m}^2 \text{ s}^{-1}$ at the Equator decreasing with the reduction of the grid spacing with latitude – it becomes $< 500 \text{ m}^2 \text{ s}^{-1}$ poleward of 60° latitude. A spatially varying field of the GM eddy advection coefficient is calculated as a function of local Rossby radius and Eady eddy-growth rate (cf. Held and Larichev, 1996). Both isoneutral diffusion and the GM eddy advection are implemented with a triad formalism (Griffies et al., 1998; Griffies, 1998). Within the surface mixed-layer, lateral diffusion is along slopes linearly decreasing with depth from the isoneutral slope immediately below the mixed layer to zero (flat) at the surface. These linearly varying slopes are also used to calculate the GM skew-fluxes: this is equivalent to a GM eddy-induced velocity that is uniform through the mixed layer (Treguier et al., 1997). This approach, used in OPA since 1999 (Madec, pers. comm.), is a simplified version of the approach recommended by Danabasoglu et al. (2008).

Lateral viscosity is parameterized by a horizontal Laplacian operator with free slip boundary condition and an eddy viscosity coefficient of $2 \times 10^4 \text{ m}^2 \text{ s}^{-1}$ except in the tropics where it reduces to $1000 \text{ m}^2 \text{ s}^{-1}$ (except along western boundaries). Finally, the diffusive component of the bottom boundary layer scheme of Döscher and Beckmann (2000) is employed, in which tracers are diffused downslope, using a diffusivity of $1000 \text{ m}^2 \text{ s}^{-1}$.

Appendix B. CORE-II IAF Experimental Protocol

We summarize the protocol for conducting CORE-II IAF experiments here, with further details provided in Griffies et al. (2012).

The ocean models are initialized with zero velocities and the January-mean climatological θ and S from the Polar Science Center Hydrographic Climatology (PHC2; a blending of the Levitus et al. (1998) data set with modifications in the Arctic Ocean based on Steele et al. (2001)). More recent θ and S data sets can also be used. The sea-ice models are generally initialized from a state available from other, existing simulations. Because the CORE-II IAF experiments are run no less than 300 years, fine details of the initial conditions are not crucial.

The surface fluxes of heat, freshwater / salt, and momentum are determined using the CORE-II IAF atmospheric data sets, the model’s prognostic SST and surface currents, and the bulk formulae described in Large and Yeager (2004, 2009). As the forcing data-sets have been developed using the formulae described in these references, we recommend using the same bulk formulae. There is no restoring term applied to SSTs. In contrast, a form of sea surface salinity (SSS) restoring may be used to prevent unbounded local salinity trends (see Appendix C for details of SSS restoring used by the groups). This restoring can be applied as either a salt flux or a *converted* water flux – the latter is for models that employ fresh water fluxes. However, the former method is preferred even for models that employ fresh water fluxes to maintain simple diagnostic control over the total water budget without any confusion from water fluxes from restoring. A modified version of the PHC2 monthly-mean SSS climatology which includes salinity enhancements along the Antarctic coast due to Doney and Hecht (2002) is recommended as the restoring field.

In contrast with the river runoff data used in Griffies et al. (2009), the new

runoff data are not pre-spread. Therefore, the user must choose how to insert river water into the ocean. For example, in AWI, the runoff flux is distributed around the river mouths with a linear function within 400 km distance. In NCAR, river runoff is spread substantially prior to applying it as a flux into the uppermost grid cell with a newer smoothing algorithm than was used in Large and Yeager (2004), yielding far less spreading. GFDL-MOM simulations choose to apply two passes of a Laplacian (1-2-1) filter in the horizontal at each time step to spread the river runoff outward from the river insertion point, resulting in a rather small spread. In addition, as detailed in Griffies et al. (2005), river runoff is inserted to the GFDL-MOM simulations over the upper four grid cells (roughly 40 m). This insertion is meant to *parameterize* tidal mixing near river mouths, and it may serve a similar purpose to the horizontal spreading applied by NCAR. In so doing, it helps to mix the fresh water throughout the upper four model grid cells, thus reducing the tendency for the simulation to produce a highly stratified fresh cap at the river mouths.

The ocean – sea-ice coupled model is run for no less than 5 repeating cycles of the 60-year forcing. Upon reaching the end of 2007, the forcing is returned to 1948. Analysis of the ocean fields during the 5th cycle provides the basis for comparing to other simulations. We note that the 60-year repeat cycling introduces an unphysical jump in the forcing from 2007 back to 1948 with the ocean state in 1948 identical to that of the end state of the previous cycle. Nevertheless, no agreeable alternative has been proposed and tested.

To aid in assessing the models’ mixing processes, ventilation rates, deep water formation, and circulation characteristics under CORE-II IAF forcing, we recommend that the simulations include ideal age tracer and chlorofluorocarbons (CFCs). The ideal age tracer (Thiele and Sarmiento, 1990) is set to zero in the model surface layer (level) at each time step, and ages at 1 year per year below. It evolves accord-

1354 ing to the same advection - diffusion equation in the ocean interior just as a passive
1355 tracer. Regions of low ventilation have the oldest waters while the younger waters
1356 indicate recent contact with the ocean surface. For a proper comparison of model
1357 ideal age distributions, we recommend that the ideal age be initialized with zero at
1358 the beginning of the 300-year simulations (five forcing cycles).

1359 The CFC-11 and CFC-12 have been increasingly utilized in evaluating ocean
1360 models largely due to i) a good observational data base (the World Ocean Circulation
1361 Experiment, WOCE, upon which Global Ocean Data Analysis Project, GLODAP,
1362 Key et al. (2004) is largely based), ii) their well-known atmospheric concentrations,
1363 and iii) because they are inert in the ocean. The surface concentrations of CFC-12
1364 and CFC-11 are available starting from 1931 and 1938, respectively. The associated
1365 fluxes should be calculated following the Ocean Carbon Model Inter-comparison
1366 Project (OCMIP-2) protocols (Dutay et al., 2002). However, instead of the protocol
1367 specified fields, the CORE-II IAF data sets should be used in the flux equations.

1368 There is a mismatch between the CFC and CORE-II IAF data start dates. At
1369 NCAR, the following approach is used. Assuming a 300-year simulation, we introduce
1370 the CFC-12 and CFC-11 surface fluxes at the beginning of model years 224 and 231,
1371 respectively, in the fourth forcing cycle. Both CFCs are initialized with zero. These
1372 model years correspond to calendar years 1991 and 1998, respectively, for the surface
1373 fluxes of heat, salt, and momentum in the IAF cycle, while they correspond to
1374 calendar year 1931 for CFC-12 and calendar year 1938 for CFC-11 surface fluxes.
1375 However, by the beginning of the fifth cycle corresponding to model year 241 and
1376 calendar year 1948, all surface fluxes become synchronous, i.e., the calendar years
1377 for the atmospheric data used in all surface flux calculations are the same during the
1378 fifth cycle. Another option is to simply introduce both CFCs at the beginning of the
1379 fifth cycle, i.e., in year 1948. Because CFC concentrations are rather small during

1380 the previous years, this represents a reasonable approach.

1381 **Appendix C. Hydrological Forcing and Salinity Restoring**

1382 As discussed in Griffies et al. (2009), the ocean – sea-ice coupled systems lack
1383 many of the feedbacks present in a fully coupled framework due to the absence of
1384 an active atmospheric component. In addition, the lack of any appreciable local
1385 feedbacks between SSS and freshwater fluxes can lead to unbounded local salinity
1386 trends that can occur in response to inaccuracies in precipitation. These two factors
1387 necessitate restoring (or relaxation) of model SSS (SSS_{model}) to an observed clima-
1388 tology (SSS_{data}) in ocean – sea-ice coupled simulations. The CORE-II IAF protocol
1389 described in Appendix B does not specify a particular recipe for such restoring and
1390 it is left to the modelers to choose their *optimal* restoring procedure.

1391 Such SSS restoring remains part of the art, rather than the science, of ocean –
1392 sea-ice climate modeling. SSS restoring is applied using a restoring salt flux of

$$1393 \quad F = V_{\text{piston}} (SSS_{\text{data}} - SSS_{\text{model}}) = V_{\text{piston}} \Delta SSS \quad (\text{C.1})$$

1394 to the top ocean model grid cell. For example, when SSS_{model} is smaller than SSS_{data} ,
1395 then a positive restoring salt flux is added. Unfortunately, the model solutions ex-
1396 hibit substantial sensitivities to the strength of the piston velocity (V_{piston}) – or
1397 equivalently to the magnitude of the restoring time scale for a given length scale,
1398 e.g., Behrens et al. (2013). It is highly desirable that the selection of a restoring time
1399 scale for a particular model is based on quantitative measures, involving compar-
1400 isons of model solutions with available observations. Often times, this decision also
1401 incorporates subjective calls involving, for example, judgments on unknown AMOC
1402 variability or making sure that the model produces a stable AMOC.

1403 An example of the sensitivity of the model AMOC simulations to the restoring
 1404 time scale is provided in Fig. 21. The figure shows several annual-mean AMOC
 1405 maximum transport time series at 26.5°N from a preliminary version of the NCAR
 1406 model in comparison with the RAPID data (Cunningham et al., 2007). The model
 1407 time series are obtained using different SSS restoring time scales: 30 days (30D); 1
 1408 year (1Y); 4 years (4Y); and infinity (NO), i.e., no restoring, all with respect to a
 1409 50 m length scale. The restoring time scale has a substantial influence on the mean
 1410 AMOC maximum transport which increases monotonically with weaker restoring
 1411 from 14.1 Sv in 30D to 20.9 Sv in NO – both over the 60-year period. Not surpris-
 1412 ingly, weaker restoring leads to larger salinity, and hence density, biases compared to
 1413 observations in the model deep water formation regions in the northern North At-
 1414 lantic (not shown). Despite these differences in the AMOC mean at this latitude, the
 1415 restoring time scale does not appear to impact the characteristics of AMOC inter-
 1416 annual to decadal variability appreciably (see also Behrens et al. (2013)). The 4Y
 1417 simulation fortuitously matches the RAPID data. We note that this metric by itself
 1418 is not sufficient to justify using a 4-year restoring time scale and additional metrics,
 1419 such as northward heat transport, θ and S differences from observations, and the
 1420 ACC transport at Drake Passage, should be considered. During this exploratory in-
 1421 vestigation, we re-confirmed that the 4-year SSS restoring time scale that has been in
 1422 use at NCAR since Large et al. (1997) produces solutions that, in general, compare
 1423 more favorably with observations than the ones obtained with the other restoring
 1424 time scales.

1425 We present a summary of the surface hydrological forcing and SSS restoring
 1426 details used by the participating groups in Table 2. Most of the groups apply real
 1427 fresh water fluxes instead of a virtual salt flux. The NEMO-based models convert
 1428 SSS restoring to a fresh water flux. All the other models apply SSS restoring as a

1429 salt flux. The restoring time scales vary considerably between the groups, but they
1430 can be gathered into three categories as follows:

- 1431 — weak restoring with time scales of about 4 years: FSU, GISS, KIEL, MIT, and
1432 NCAR,
- 1433 — moderate restoring with time scales of 9 - 12 months: AWI, BERGEN, CER-
1434 FACS, CMCC, CNRM, GFDL-MOM, ICTP, INMOM, MRI-A, and MRI-F,
1435 NOCS,
- 1436 — strong restoring with time scales of 50 - 150 days: ACCESS, GFDL-GOLD.

1437 In all models, the SSS restoring is applied globally and under ice covered regions
1438 — the latter with the exception of ICTP and KIEL. However, in ten of the models,
1439 the mismatch between SSS_{model} and SSS_{data} is limited to 0.5 psu, i.e., $|\Delta SSS| \leq 0.5$
1440 psu, to avoid extremely large salt fluxes of either sign that may occur, for example,
1441 in the vicinity of western boundary currents that are not realistically represented in
1442 coarse resolution simulations. The main idea is to minimize any spurious weakening
1443 of AMOC due to possible northward transport of too much fresh water that can be
1444 added to the model without such a limit. Some groups which use narrow river runoff
1445 spreading also choose to eliminate restoring at grid cells receiving river runoff so that
1446 freshening due to runoff would not be compensated by overly salty values found in
1447 the restoring field.

1448 To ensure that there is no accumulation of salt due to the restoring fluxes, most
1449 of the groups remove the globally integrated salt content arising from restoring at
1450 each time step. We note that this is a global correction, impacting the magnitude
1451 and even the sign of local restoring fluxes.

1452 Finally, given the evolving model SSTs, there is no guarantee that precipitation
 1453 (P) plus runoff (R) minus evaporation (E) will balance to zero so that the ocean
 1454 – sea-ice total water content – or salt content for those models using virtual salt
 1455 fluxes – will not change. All groups use some sort of normalization to enforce such
 1456 a constraint. These normalizations impact the surface ocean globally; they are non-
 1457 local. Examples include i) multiplication of P+R by a factor based on the global
 1458 salinity change in the ocean over the previous year to bring the salinity change
 1459 towards zero as in NCAR (Large et al., 1997), and ii) enforcing globally integrated
 1460 $P+R-E=0$ at each time step as in, e.g., CMCC, GFDL-MOM, GFDL-GOLD, and
 1461 MIT. Operationally, in CMCC, GFDL-MOM, and GFDL-GOLD, the global mean of
 1462 $P-E+R$ is subtracted from $P-E$; the runoff is not modified. So in effect, the global
 1463 area integrated $P-E$ will be equal and opposite in sign to the global area integrated
 1464 runoff. Additionally, water can be exchanged with the sea-ice, yet this exchange is
 1465 not considered for purposes of the global normalization used in these models.

1466 **Appendix D. Calculations of Zonal Averages and Transports**

1467 In this Appendix, we briefly summarize how the zonal averages and transports (or
 1468 integrals) are computed by the participating groups. The latter concerns calculations
 1469 of AMOC and MHT.

1470 Due to its regular longitude - latitude grid, GISS is the only model that does
 1471 not require any additional regridding to obtain true zonal averages and integrals. In
 1472 AWI, FSU, INMOM, MRI-A, and MRI-F, variables are first interpolated to regular
 1473 longitude - latitude grids and then zonal operations are performed. In NCAR, a
 1474 binning approach is used for transports where horizontal divergences of volume and
 1475 heat calculated on the model grid are summed within specified latitude bands onto
 1476 a regular longitude - latitude grid. Zonal averages in NCAR are computed using a

1477 volume-weighted average (or horizontal area-weighted average because the vertical
1478 thicknesses are the same for a given vertical level) of a field where the average is over
1479 the model grid cells intersecting the latitude band, and the horizontal area for the
1480 weighting is the area of intersection of the model grid cell with the latitude band.

1481 In ACCESS, GFDL-GOLD, GFDL-MOM, and ICTP, the model grids are truly
1482 zonal south of 65°N. Similarly, the model grids are truly zonal south of 38.5° and
1483 56°N in BERGEN and MIT, respectively. Thus, these models do not necessitate
1484 any regridding south of these latitudes. Further north, the zonal operations are
1485 performed along model grid lines, despite their deviations from constant latitude
1486 lines.

1487 All zonal calculations are done along the distorted grid lines in NEMO-based
1488 models, i.e., CERFACS, CMCC, CNRM, KIEL, and NOCS. The grid distortion is
1489 rather small at low latitudes. For example, latitude varies by about 0.03° along a
1490 model grid line (a line of constant model latitude index) near 26.9°N in the Atlantic
1491 Basin. However, the distortion gradually increases to $> 2^\circ$ by about 60°N, e.g.,
1492 the minimum and maximum latitudes are 60.1° and 62.5°N along a grid line. The
1493 nominal latitude is specified as the maximum latitude along a grid line. North of
1494 60°N, the grid distortions become larger, making such *zonal* averages and transports
1495 less meaningful.

1496 In the vertical, BERGEN, FSU, GFDL-GOLD, and INMOM use regridding or
1497 binning to map from model vertical coordinates to depth levels. In GISS, the zonal
1498 operations are done in mass levels as their depths vary only slightly with time, i.e.,
1499 by $< 1\%$.

1500 We believe that calculations of *zonal* integrations and averages along model grid
1501 lines are acceptable for our present purposes because serious grid distortions from
1502 true zonal averages are expected to occur only at high latitudes where AMOC and

1503 MHT are relatively small. However, we note that proper calculations of AMOC and
 1504 MHT at these high latitudes are important for studies involving Arctic Ocean and
 1505 sea-ice where even small transports matter significantly.

1506 **Appendix E. Impacts of Neptune Parameterization**

1507 A comparison experiment was performed of a NEMO run identical to the NOCS
 1508 contribution except that a *simplified* version of the Neptune parameterization of
 1509 unresolved eddy – topographic interactions (Eby and Holloway, 1994) was used,
 1510 following Holloway and Wang (2009). The horizontal velocity field is relaxed towards
 1511 a topographically determined, steady, Neptune velocity field

$$u^{\text{Nept}} = -\frac{1}{H} \frac{\partial \psi^{\text{Nept}}}{\partial y}, \quad v^{\text{Nept}} = \frac{1}{H} \frac{\partial \psi^{\text{Nept}}}{\partial x}, \quad (\text{E.1})$$

1512 derived from a transport stream function

$$\psi^{\text{Nept}} = -fL^2H. \quad (\text{E.2})$$

1513 Here H is the ocean depth, f the Coriolis parameter, and L is a latitude-dependent
 1514 length that scales smoothly from 12 km at the equator down to 3 km at the poles.
 1515 To avoid excessively strong flow in shallow waters, \mathbf{u}^{Nept} is scaled linearly down to
 1516 zero as H shallows from 200 to 100 m.

1517 The resulting ψ^{Nept} is quite significant, ranging from -27 Sv at 30°N to -13.9 Sv
 1518 at 60°N and -5.2 Sv at the North Pole, and has a major impact on the solutions in
 1519 NOCS. Specifically, the cyclonic topographic flow thus excited in the Greenland Sea,
 1520 LS, and subpolar gyre brings down cool fresh water (and ice) from the Greenland
 1521 current into the LS. This quenches winter convection in the LS, reducing winter

MLDs to 100–200 m, and freshens and cools the whole subpolar gyre. This freshening and cooling even penetrates into the western subtropical gyre.

This Neptune experiment has a much weaker MHT, with a maximum of only 0.42 PW compared with the 0.69 PW in the standard NOCS contribution. However, the maximum AMOC differs little. Figure 22a shows the differences (Neptune – standard) in temperature flux along a quasi-zonal section at 26.5°N, near where the maximum MHT is achieved, for the last year of the run. These differences are similar to the annual-mean differences in Fig. 22b of v (meridional velocity) times the average of the temperatures between the Neptune and standard experiments, implying that they are driven by changes in the flow rather than in the temperature. This suggests that Neptune reduces the integrated MHT because the cyclonic boundary current weakens northward transport in the Gulf Stream, where the water is warm, and strengthens northward transport along the eastern edge, where the water is cool.

Indeed, the plots of the AMOC in Fig. 22c show that the overturning circulation differs little, but the plot of the difference in the cumulative vertically integrated heat transport (Fig. 22d) shows how the weaker zonally integrated heat transports with Neptune in the upper 1000 m reduce the total MHT by about 0.3 PW. Again, this cumulative heat flux difference (blue line) is largely due to changes in the velocity field (green line).

Our experience with the Neptune parameterization appears to be consistent with that of Roubicek et al. (1995). Despite some improvements of the mid-latitude jet separation location, they find that the strong cyclonic circulation produced by the parameterization dominates the barotropic circulation in idealized, wind-driven experiments with large topographic slopes. Using a biharmonic implementation of the Neptune parameterization in an eddying global model, Maltrud and Holloway (2008) report only marginal improvements in the Gulf Stream and Arctic Ocean solutions

1548 with no obvious degradations elsewhere. In contrast, Holloway and Wang (2009)
 1549 (see also Holloway et al., 2007) show improvements of the Arctic Ocean solutions
 1550 with this parameterization in comparison with those obtained with frictional param-
 1551 eterizations. In light of these mixed results, we concur with the above studies in
 1552 their suggestions for refinements of the Neptune parameterization for both coarse
 1553 and eddy permitting / resolving applications.

1554 **Appendix F. List of Major Acronyms**

- 1555 — ACCESS: Australian Community Climate and Earth System Simulator
- 1556 — AWI: Alfred Wegener Institute
- 1557 — CCSM: Community Climate System Model
- 1558 — CERFACS: Centre Européen de Recherche et de Formation Avancée en Calcul
 1559 Scientifique
- 1560 — CESM: Community Earth System Model
- 1561 — CICE: Sea ice model
- 1562 — CLIVAR: Climate Variability and Predictability
- 1563 — CM: Coupled model
- 1564 — CMCC: Centro Euro-Mediterraneo sui Cambiamenti Climatici
- 1565 — CMIP5: Coupled Model Inter-comparison Project phase 5
- 1566 — CNRM: Centre National de Recherches Météorologiques
- 1567 — CORE: Coordinated Ocean-ice Reference Experiments

- 1568 – CSIM: Community Sea Ice Model
- 1569 – CSIRO: Commonwealth Scientific and Industrial Research Organisation
- 1570 – DRAKKAR: Coordination of high resolution global ocean simulations and de-
- 1571 velopments of the NEMO modelling framework
- 1572 – ECCO: Estimating the Circulation and Climate of the Ocean
- 1573 – EVP: Elastic-viscous-plastic
- 1574 – FESOM: Finite Element Sea-ice Ocean Model
- 1575 – FSU: Florida State University
- 1576 – GFDL: Geophysical Fluid Dynamics Laboratory
- 1577 – GISS: Goddard Institute for Space Studies
- 1578 – GM: Gent and McWilliams (1990)
- 1579 – GOLD: Generalized Ocean Layer Dynamics
- 1580 – HYCOM: HYbrid Coordinate Ocean Model
- 1581 – IAF: Inter-annual forcing
- 1582 – ICTP: International Centre for Theoretical Physics
- 1583 – INMCM: Institute of Numerical Mathematics Earth Climate Model
- 1584 – INMOM: Institute of Numerical Mathematics Ocean Model
- 1585 – KPP: K-Profile Parameterization (Large et al., 1994)

- 1586 — LANL: Los Alamos National Laboratory
- 1587 — LIM: Louvain-la-Neuve Sea Ice Model
- 1588 — LS: Labrador Sea
- 1589 — MICOM: Miami Isopycnal Coordinate Ocean Model
- 1590 — MIT: Massachusetts Institute of Technology
- 1591 — MITgcm: Massachusetts Institute of Technology general circulation model
- 1592 — MOM: Modular Ocean Model
- 1593 — MOVE: Multivariate Ocean Variational Estimation
- 1594 — MRI: Meteorological Research Institute
- 1595 — MRI.COM: Meteorological Research Institute Community Ocean Model
- 1596 — NASA: National Aeronautics and Space Administration
- 1597 — NCAR: National Center for Atmospheric Research
- 1598 — NEMO: Nucleus for European Modelling of the Ocean
- 1599 — NOAA: National Oceanic and Atmospheric Administration
- 1600 — NOCS: National Oceanography Centre Southampton
- 1601 — NorESM-O: Norwegian Earth System Model ocean component
- 1602 — NYF: Normal-year forcing
- 1603 — OASIS: A European coupling framework for components of the climate system

- 1604 — OPA: Ocean PArallelise, the Ocean General Circulation Model developed at
1605 the Laboratoire d’Oceanographie DYnamiquexi et de Climatologie (LODYC).
- 1606 — ORCA: Ocean model configuration of the NEMO model
- 1607 — PHC: Polar Science Center Hydrographic Climatology
- 1608 — POP2: Parallel Ocean Program version 2
- 1609 — SIS: GFDL Sea Ice Simulator
- 1610 — SPG: Subpolar gyre
- 1611 — STG: Subtropical gyre
- 1612 — WGOMD: Working Group on Ocean Model Development
- 1613 — WOA: World Ocean Atlas

1614 References

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2209 List of Figures

2210	1	AMOC annual-mean maximum transport time series at 26.5°N for	
2211		the entire 300-year integration length. The time series are smoothed	
2212		using a five-point box car filter. The repeating 60-year forcing cycle,	
2213		corresponding to calendar years 1948-2007, is indicated by the dashed	
2214		lines in each panel.	102
2215	2	Root-mean-square (rms) differences (top panels) and correlations (bot-	
2216		tom panels) for the AMOC annual-mean maximum transport time se-	
2217		ries at 26.5°N between consecutive forcing cycles. The first ten years	
2218		of each cycle are excluded from the analysis to avoid large adjust-	
2219		ments associated with the jump in forcing from 2007 back to 1948.	
2220		The MRI-A data assimilation simulation is not included because it is	
2221		integrated only for one 60-year cycle.	103
2222	3	Time-mean AMOC plotted in depth (km) and latitude space. The	
2223		positive and negative contours indicate clockwise and counter-clockwise	
2224		circulations, respectively. In MIT, AWI, MRI-F, MRI-A, FSU, BERGEN,	
2225		and GISS, the AMOC distributions do not include the high latitude	
2226		North Atlantic and / or Arctic Oceans, and hence are masked. Unless	
2227		otherwise noted, the time-mean refers to the 20-year means for years	
2228		1988-2007, corresponding to simulation years 281-300, in all the figures.	104
2229	4	Same as in Fig. 3 except for AMOC plotted in σ_2 density (kg m^{-3})	
2230		and latitude space. INMOM distribution is not available.	105
2231	5	Years 2004-2007 mean AMOC depth profiles at 26.5°N from model	
2232		solutions in comparison with the 4-year mean (April 2004 - March	
2233		2008) RAPID data (thick black lines plotted in each panel).	106

2234	6	Time-mean meridional heat transports for the Atlantic Ocean. The	
2235		black lines denoted by L&Y09 represent implied time-mean transport	
2236		calculated by Large and Yeager (2009) with shading showing the im-	
2237		plied transport range in individual years for the 1984-2006 period.	
2238		Direct estimates with their uncertainty ranges from the RAPID data	
2239		(square; Johns et al., 2011) and from Bryden and Imawaki (2001)	
2240		(triangle; B&I01) are also shown.	107
2241	7	Scatter plot of the maximum AMOC transport against meridional	
2242		heat transport (MHT), both evaluated at 26.5°N. The model data are	
2243		for the time-mean. The solid star denotes the observational AMOC	
2244		and MHT estimates from the RAPID data. The regression line and	
2245		correlation coefficient are also shown.	108
2246	8	Time-mean, 0-700 m average potential temperature model minus ob-	
2247		servations (Locarnini et al., 2010) difference distributions. The boxed	
2248		area in the NCAR panel indicates the Labrador Sea analysis region. .	109
2249	9	Time-mean, 0-700 m average salinity model minus observations (Antonov	
2250		et al., 2010) difference distributions.	110
2251	10	Time-mean, 0-700 m average density model minus observations dif-	
2252		ference distributions. The observational density is based on WOA09	
2253		temperature and salinity.	111
2254	11	Time-mean and zonal-mean potential temperature model minus ob-	
2255		servations (Locarnini et al., 2010) difference distributions for the At-	
2256		lantic Ocean. In MIT, MRI-F, KIEL, CERFACS, CNRM, MRI-A,	
2257		FSU, and GISS, the difference distributions do not include the high	
2258		latitude North Atlantic and / or Arctic Oceans, and hence are masked.	112

2259	12	Same as in Fig. 11 except for salinity, and observations are from	
2260		Antonov et al. (2010)	113
2261	13	March-mean mixed layer depth (MLD) based on a $\Delta\rho = 0.125$ kg	
2262		m ⁻³ criterion for the northern North Atlantic. The panel to the left	
2263		of the color bar shows MLD calculated from the WOA09 potential	
2264		temperature (Locarnini et al., 2010) and salinity (Antonov et al., 2010)	
2265		data.	114
2266	14	Time-mean and zonal-mean ideal age distributions for the Atlantic	
2267		Ocean. Ideal age is not available from MIT, ACCESS, ICTP, KIEL,	
2268		CERFACS, CNRM, and FSU. In MRI-F and MRI-A, the distributions	
2269		do not include the Arctic Ocean, and hence are masked.	115
2270	15	March-mean sea-ice thickness. The black contour line denotes the	
2271		15% observed sea-ice extent from the updated NSIDC data (Cavalieri	
2272		et al., 1996, updated yearly).	116
2273	16	Scatter plots of the North Atlantic subpolar gyre (SPG) maximum	
2274		transports against the North Atlantic subtropical gyre (STG) maxi-	
2275		imum transports (left) and against the overflow proxy density (right).	
2276		All data are time-mean. Overflow proxy density is not available from	
2277		INMOM. See text for details of how STG and SPG transports are de-	
2278		termined. The overflow proxy density is calculated as the time-mean	
2279		density of the densest outflow (or southward flow) at 60°N as repre-	
2280		sented by approximately 1 Sv AMOC transport in density (σ_2) space,	
2281		using Fig. 4.	117

2282	17	Scatter plots of the Labrador Sea (LS) upper-ocean (0–700 m aver-	
2283		age) potential temperature (left), salinity (middle), and density (right)	
2284		biases against the AMOC maximum transports at 26.5°N (top) and	
2285		45°N (bottom). Each panel contains the corresponding regression line	
2286		along with the correlation coefficient. The model data are for the	
2287		time-mean. The solid stars in the top panels show the 4-year mean	
2288		RAPID data (April 2004 - March 2008) plotted against no bias. The	
2289		LS region is bounded by 60°-45°W and 50°-65°N and is indicated in	
2290		Fig. 8.	118
2291	18	Scatter plots of the LS March-mean MLD against the AMOC maxi-	
2292		mum transports at 26.5°N and 45°N (top panels) and against the LS	
2293		upper-ocean (0–700 m average) potential temperature, salinity, and	
2294		density biases (bottom panels). Each panel contains the correspond-	
2295		ing regression line along with the correlation coefficient. Except MLD,	
2296		the model data are for the time-mean. The solid stars in the top left	
2297		and bottom panels represent the observationally-based March-mean	
2298		MLD estimate plotted against the 4-year mean RAPID data (April	
2299		2004 - March 2008) and against no bias, respectively.	119
2300	19	Scatter plots of the AMOC maximum transports at 26.5°N (left) and	
2301		45°N (right) against the overflow proxy density described in Fig. 16.	
2302		All data are time-mean. Overflow proxy density is not available from	
2303		INMOM.	120

2304	20	Scatter plots of the LS March-mean sea-ice area against the LS upper-	
2305		ocean (0–700 m average) potential temperature (top left), salinity	
2306		(top middle), and density (top right) biases, and against the LS March-	
2307		mean MLD (bottom left), AMOC maximum transport at 26.5°N (bot-	
2308		tom middle), and AMOC maximum transport at 45°N (bottom right).	
2309		Each panel contains the corresponding regression line along with the	
2310		correlation coefficient. Except MLD and sea-ice area, the model data	
2311		are for the time-mean. The solid stars in the top panels show observa-	
2312		tional March-mean sea-ice area plotted against no bias. The solid stars	
2313		in the lower left and middle panels are for the observational March-	
2314		mean sea-ice area plotted against the observationally-based March-	
2315		mean MLD and the 4-year mean RAPID data (April 2004 - March	
2316		2008), respectively.	121
2317	21	AMOC annual-mean maximum transport time series at 26.5°N ob-	
2318		tained with a preliminary version of the NCAR model using different	
2319		SSS restoring time scales: 30 days (30D); 1 year (1Y); 4 years (4Y);	
2320		and infinity (NO), i.e., no restoring. The associated length scale is 50	
2321		m. RAPID line represents the observational data from Cunningham	
2322		et al. (2007). The time series are shown for one forcing cycle.	122

2323 22 a) Difference (Neptune – standard) in northward annual-mean tem-
 2324 perature flux along a quasi-zonal section at 26.5°N in °C m s⁻¹. b)
 2325 Difference in velocity times mean of the temperatures from the two
 2326 runs in °C m s⁻¹. c) Zonally integrated meridional flow cumulatively
 2327 integrated up from the bottom in Sv for the Neptune run (blue line)
 2328 and standard run (green line). d) Difference in zonally integrated
 2329 heat transport cumulatively integrated up from the bottom in PW,
 2330 calculated using the full temperature flux (blue line) and the approx-
 2331 imation, using only the difference in velocities (green line). All plots
 2332 are for year 2007 of the last forcing cycle. 123

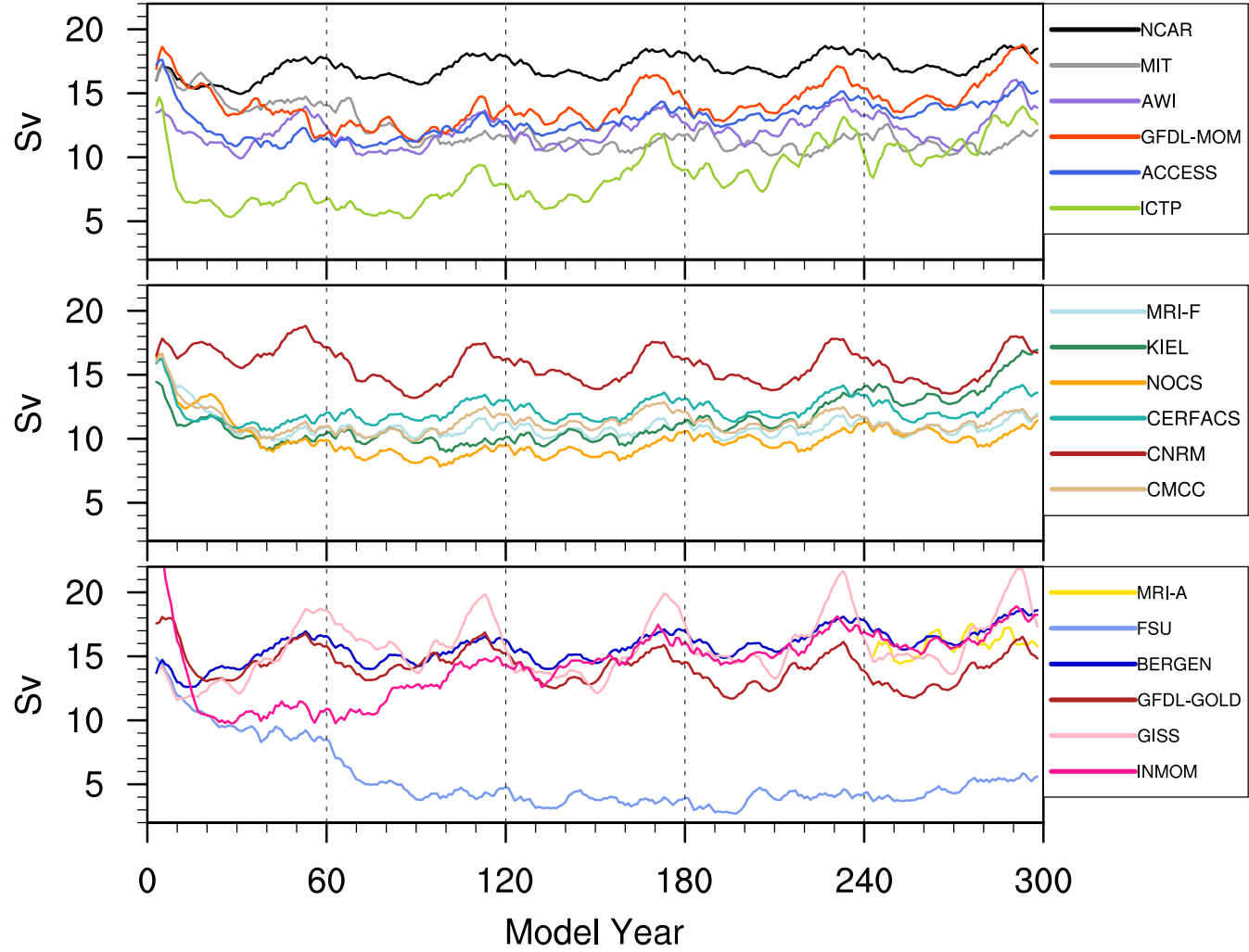


Figure 1: AMOC annual-mean maximum transport time series at 26.5°N for the entire 300-year integration length. The time series are smoothed using a five-point box car filter. The repeating 60-year forcing cycle, corresponding to calendar years 1948-2007, is indicated by the dashed lines in each panel.

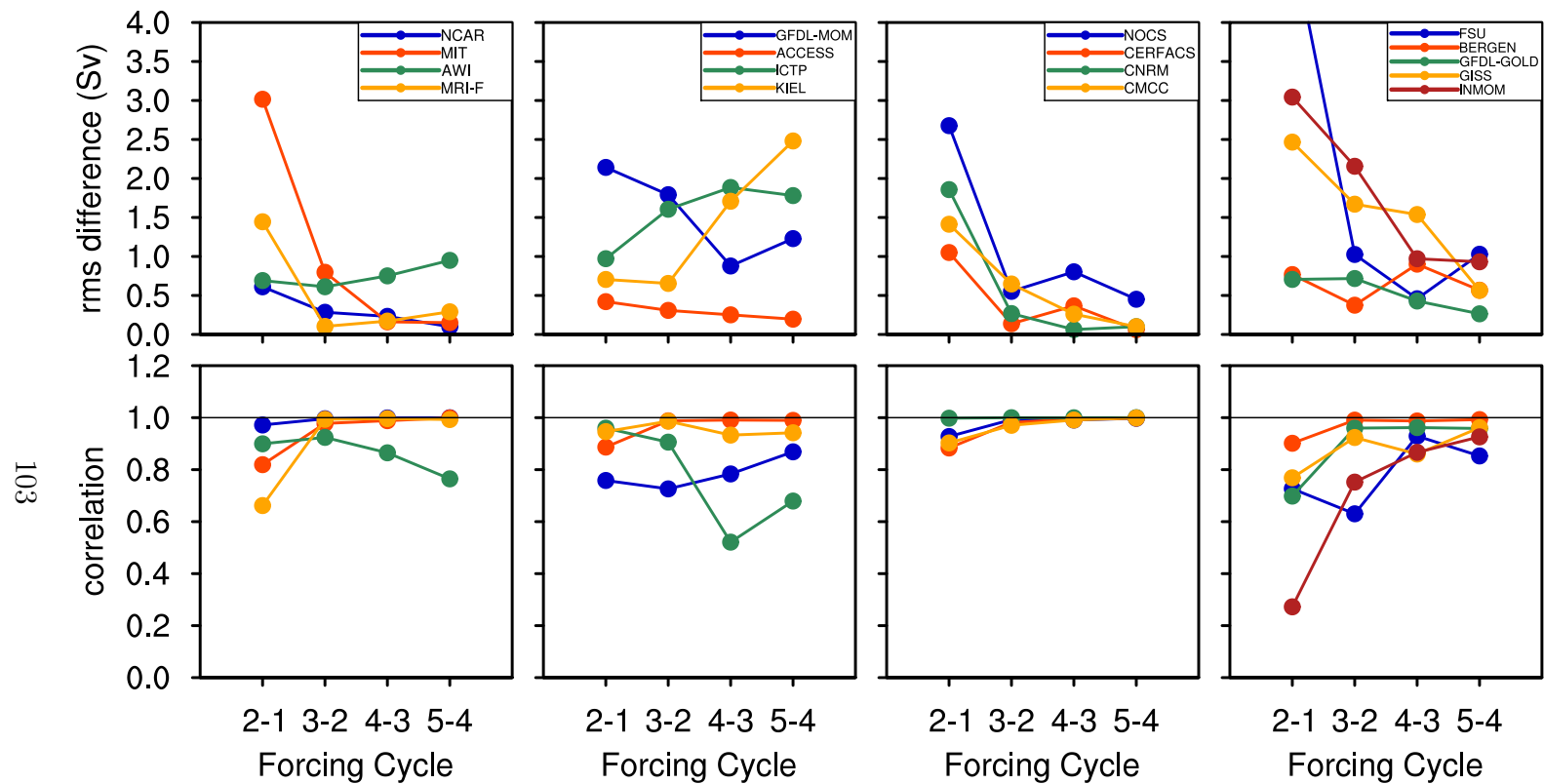


Figure 2: Root-mean-square (rms) differences (top panels) and correlations (bottom panels) for the AMOC annual-mean maximum transport time series at 26.5°N between consecutive forcing cycles. The first ten years of each cycle are excluded from the analysis to avoid large adjustments associated with the jump in forcing from 2007 back to 1948. The MRI-A data assimilation simulation is not included because it is integrated only for one 60-year cycle.

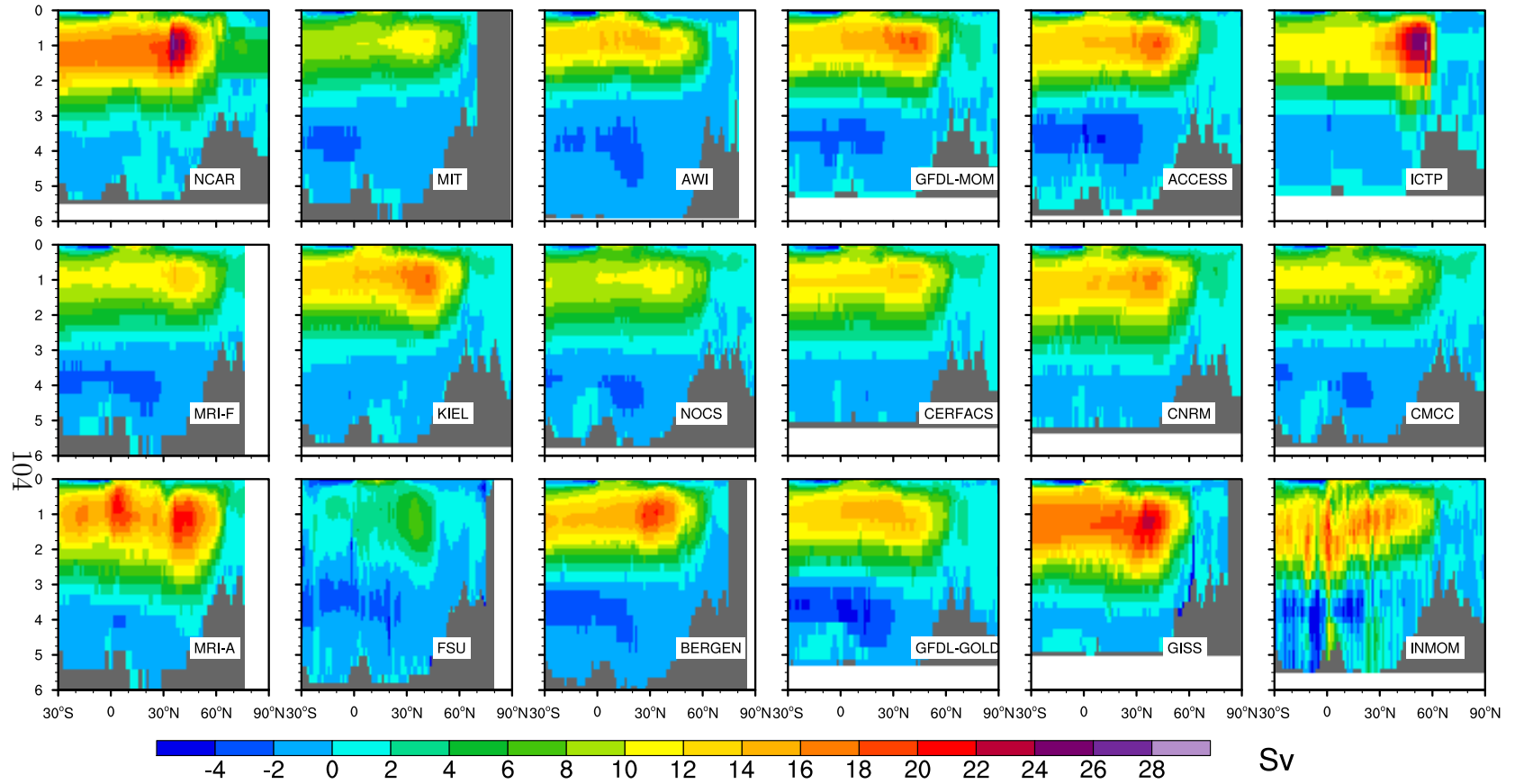


Figure 3: Time-mean AMOC plotted in depth (km) and latitude space. The positive and negative contours indicate clockwise and counter-clockwise circulations, respectively. In MIT, AWI, MRI-F, MRI-A, FSU, BERGEN, and GISS, the AMOC distributions do not include the high latitude North Atlantic and / or Arctic Oceans, and hence are masked. Unless otherwise noted, the time-mean refers to the 20-year means for years 1988-2007, corresponding to simulation years 281-300, in all the figures.

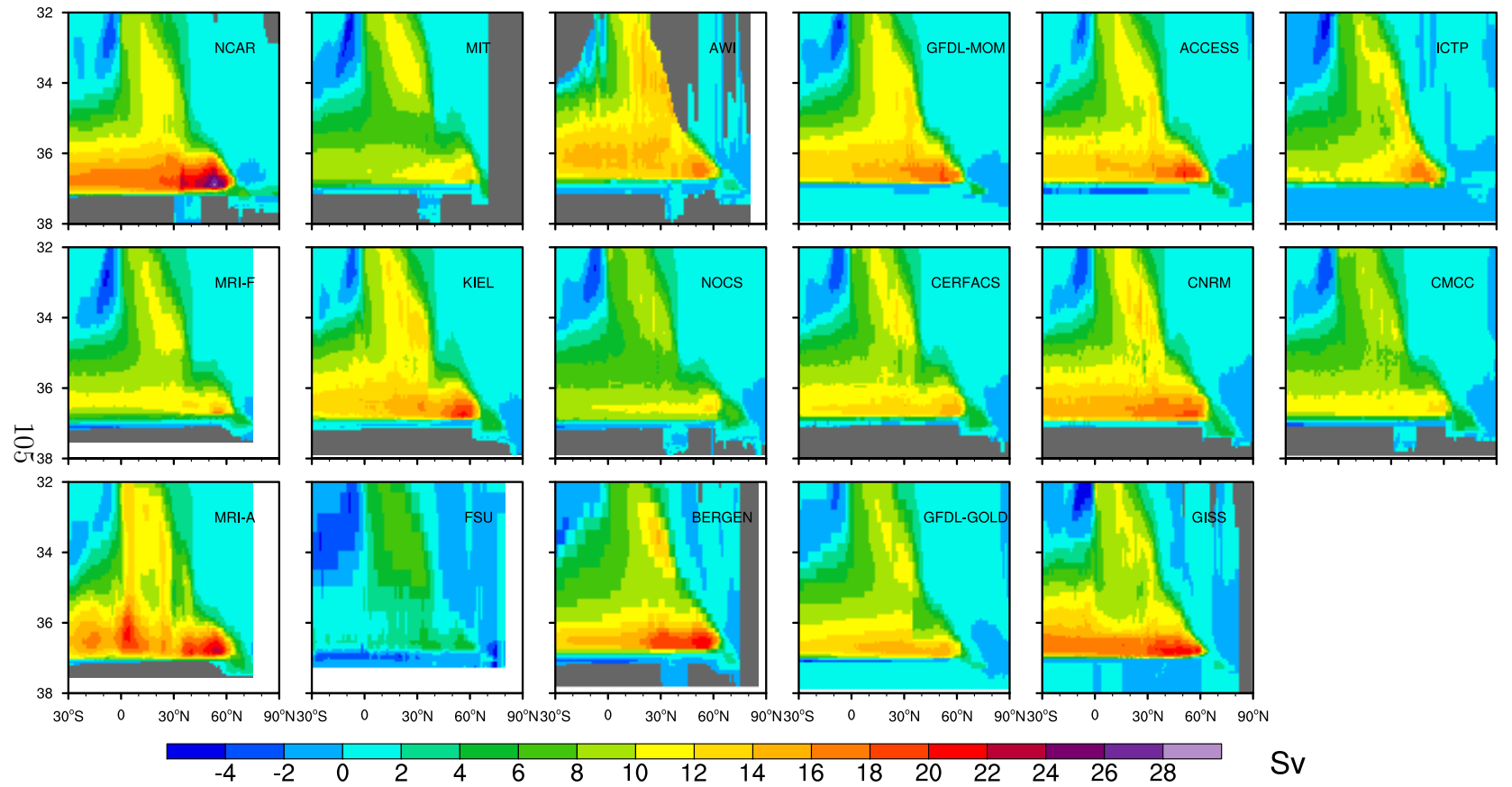


Figure 4: Same as in Fig. 3 except for AMOC plotted in σ_2 density (kg m^{-3}) and latitude space. INMOM distribution is not available.

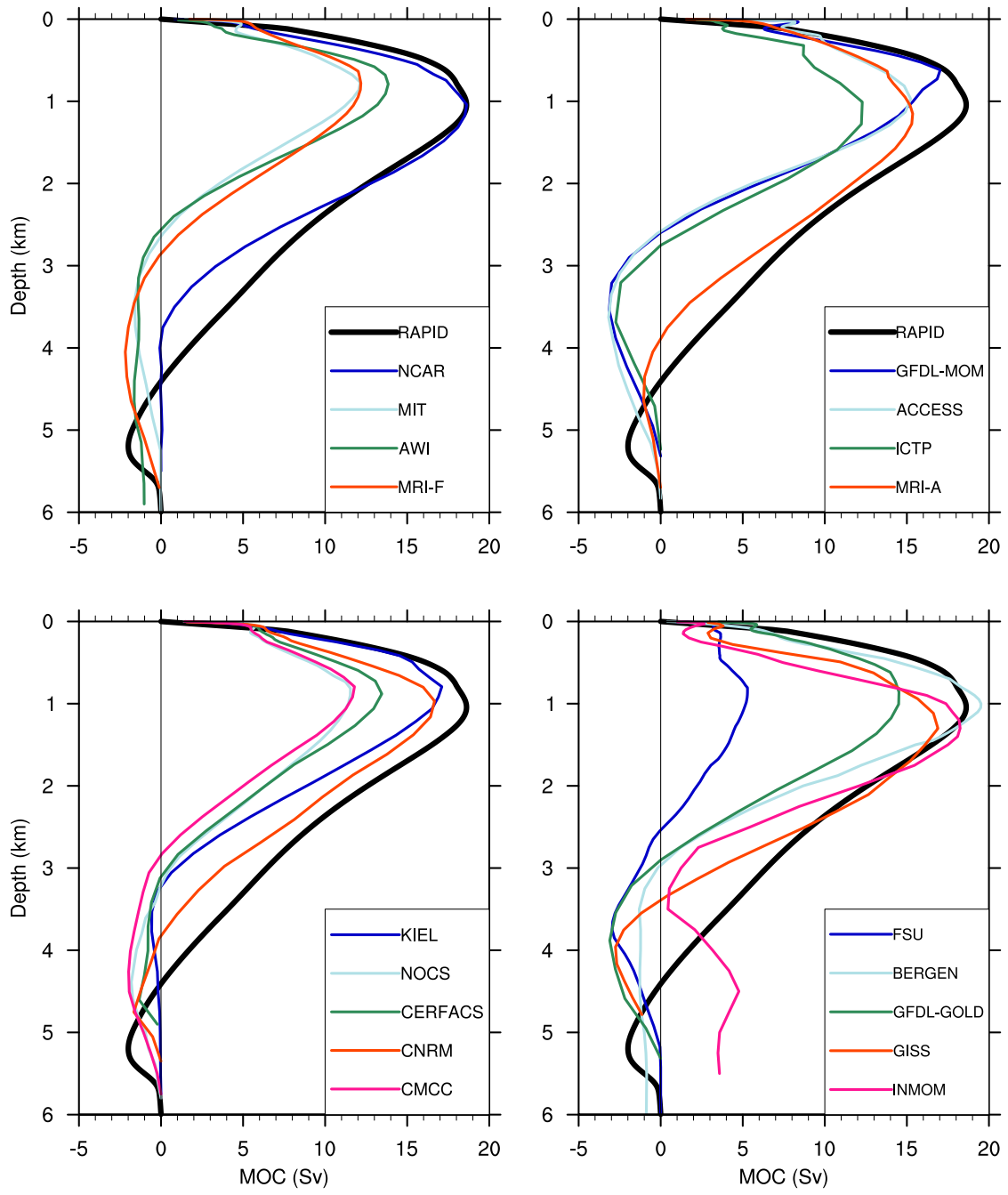


Figure 5: Years 2004-2007 mean AMOC depth profiles at 26.5°N from model solutions in comparison with the 4-year mean (April 2004 - March 2008) RAPID data (thick black lines plotted in each panel).

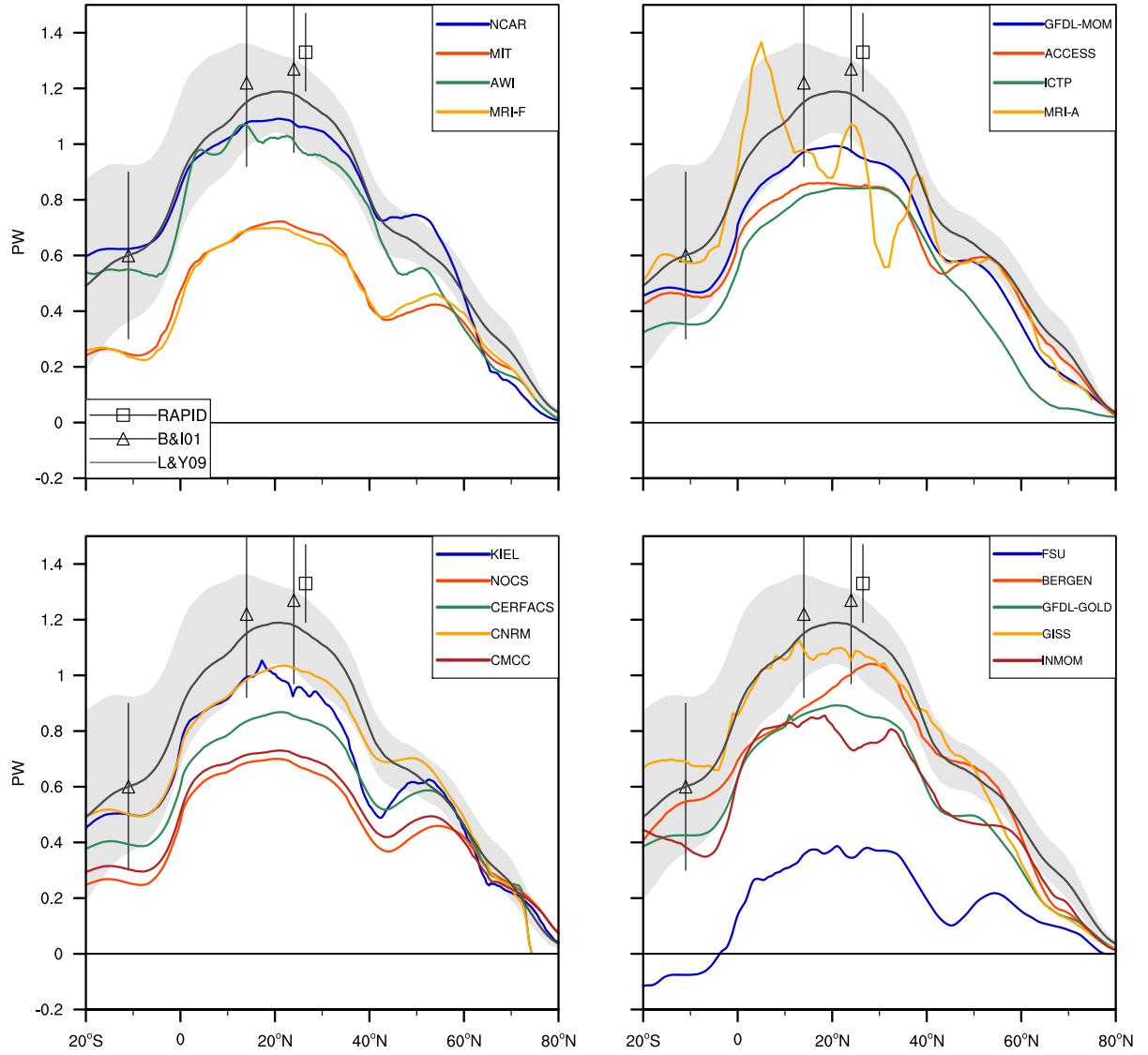


Figure 6: Time-mean meridional heat transports for the Atlantic Ocean. The black lines denoted by L&Y09 represent implied time-mean transport calculated by Large and Yeager (2009) with shading showing the implied transport range in individual years for the 1984-2006 period. Direct estimates with their uncertainty ranges from the RAPID data (square; Johns et al., 2011) and from Bryden and Imawaki (2001) (triangle; B&I01) are also shown.

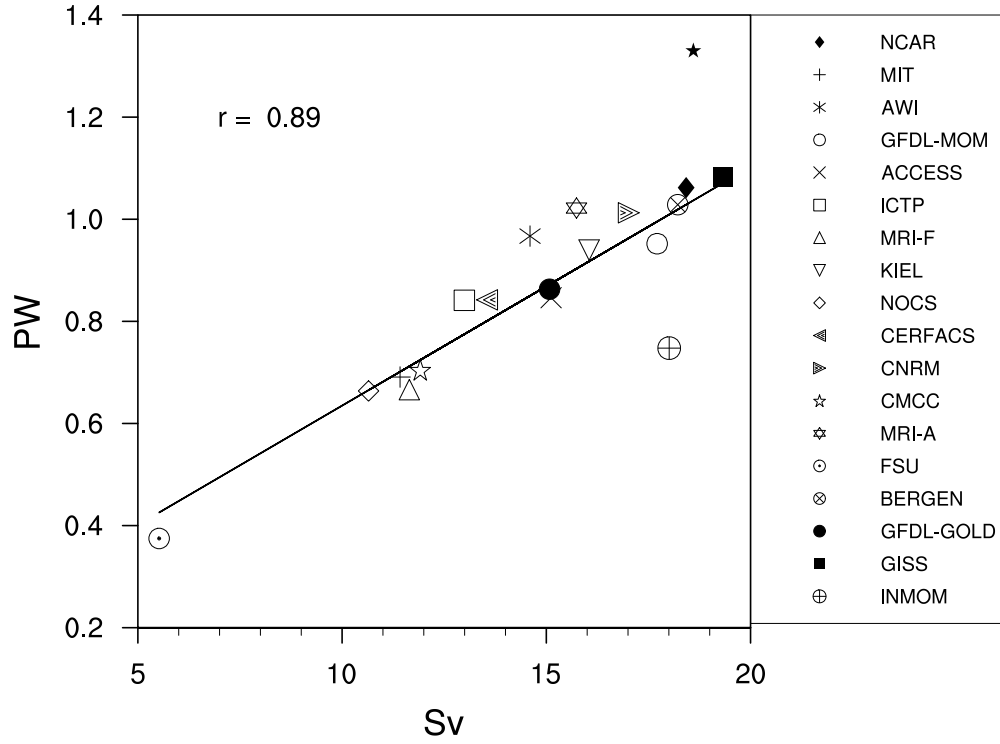


Figure 7: Scatter plot of the maximum AMOC transport against meridional heat transport (MHT), both evaluated at 26.5°N. The model data are for the time-mean. The solid star denotes the observational AMOC and MHT estimates from the RAPID data. The regression line and correlation coefficient are also shown.

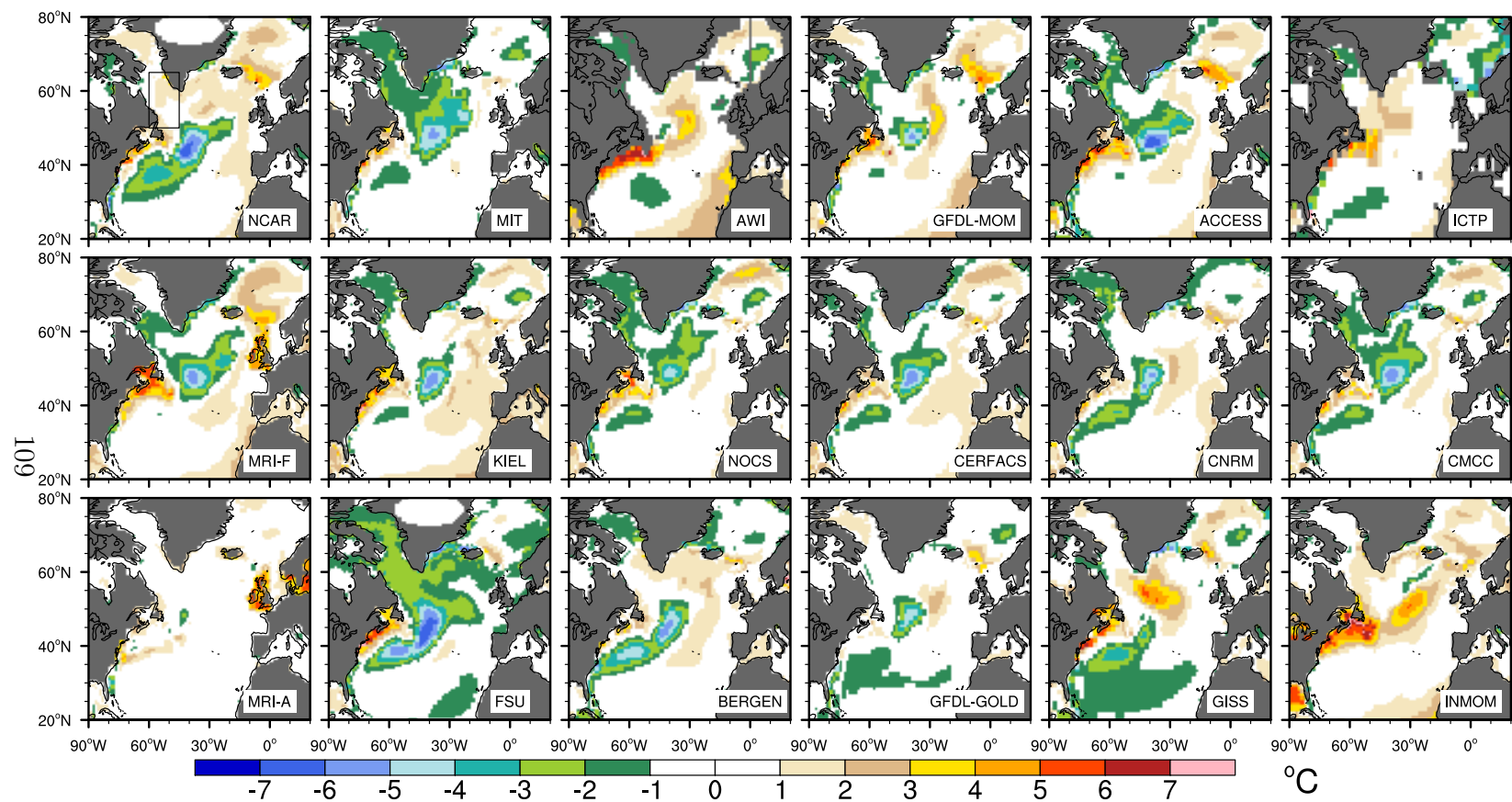


Figure 8: Time-mean, 0-700 m average potential temperature model minus observations (Locarnini et al., 2010) difference distributions. The boxed area in the NCAR panel indicates the Labrador Sea analysis region.

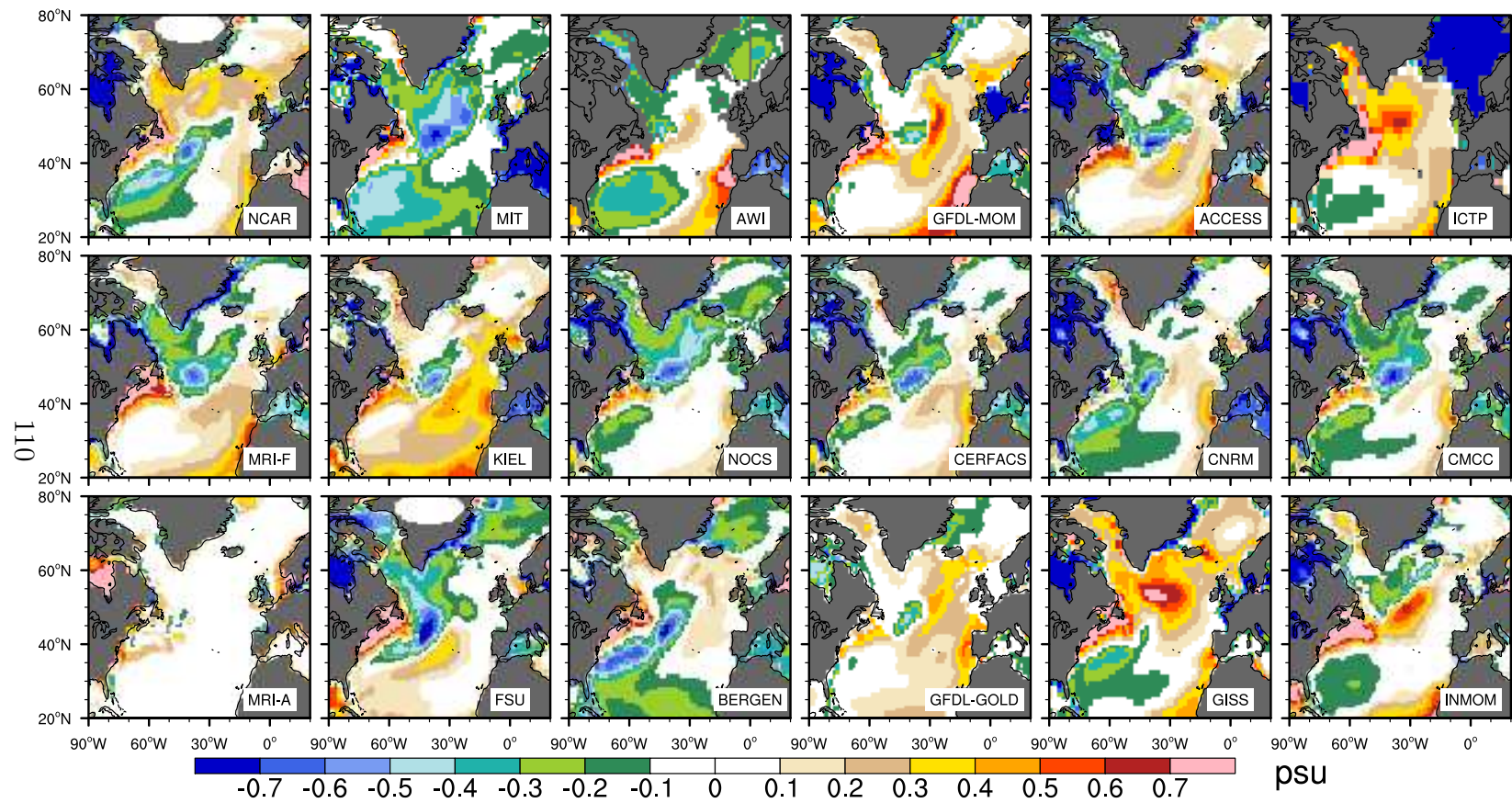


Figure 9: Time-mean, 0-700 m average salinity model minus observations (Antonov et al., 2010) difference distributions.

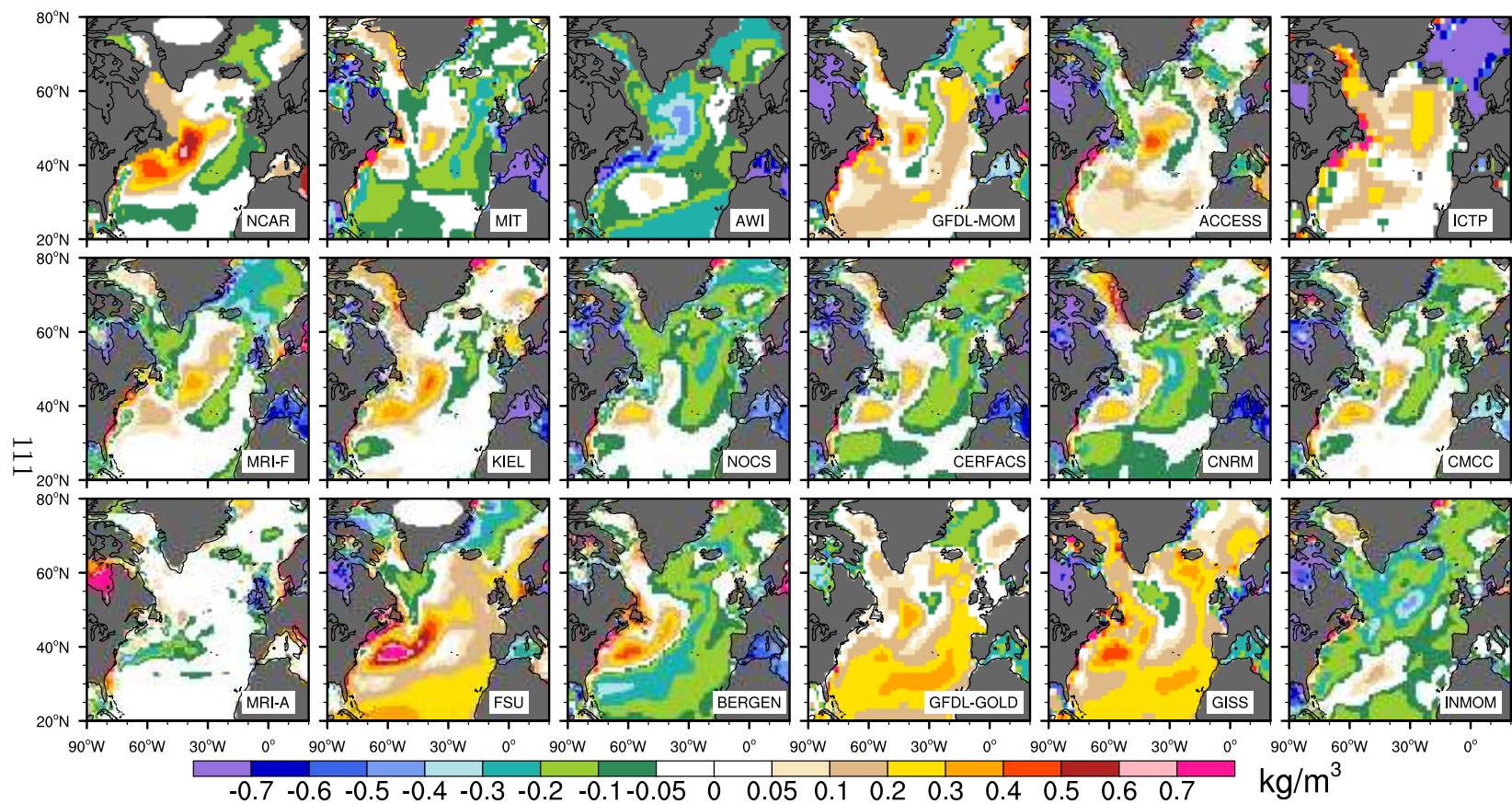


Figure 10: Time-mean, 0-700 m average density model minus observations difference distributions. The observational density is based on WOA09 temperature and salinity.

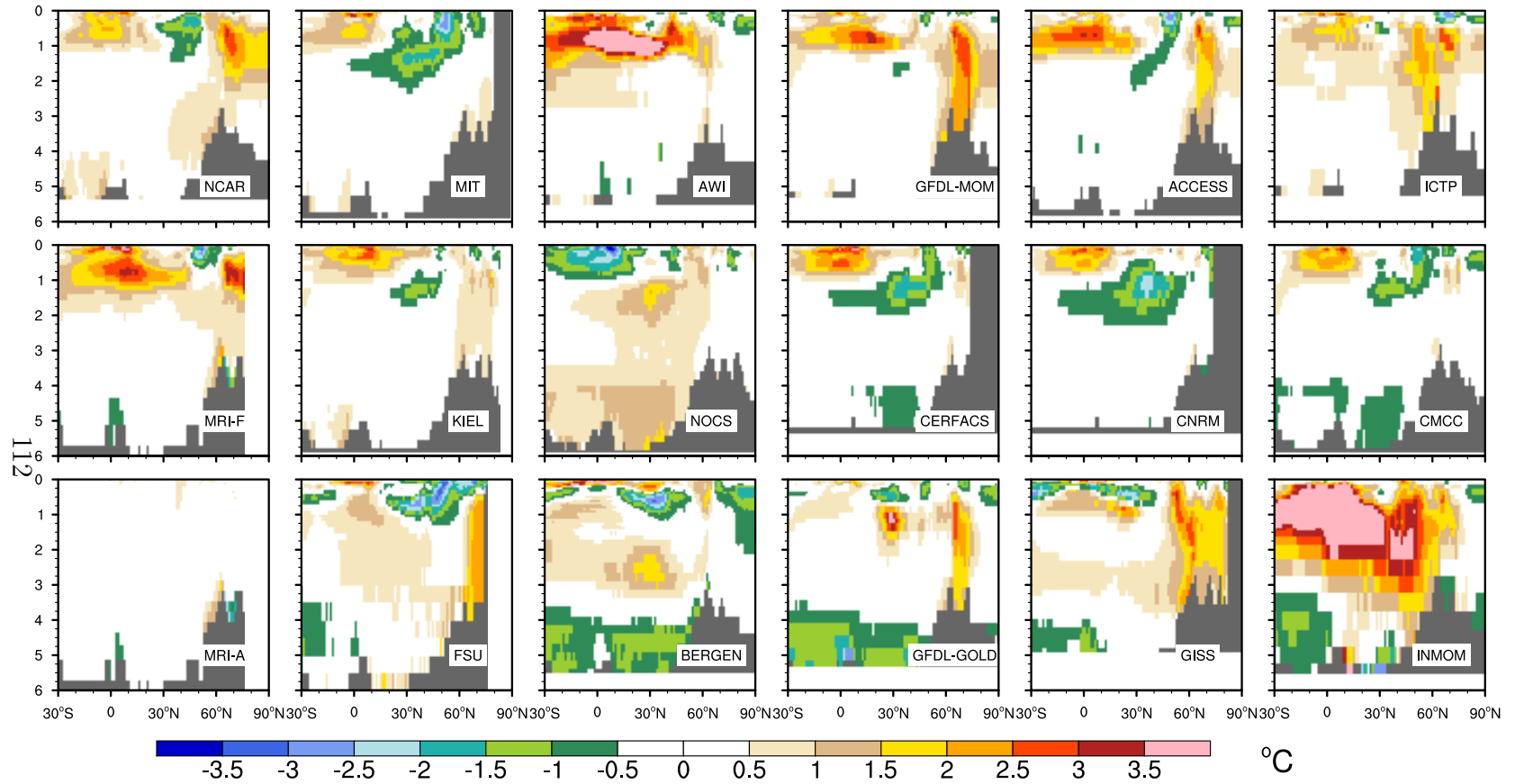


Figure 11: Time-mean and zonal-mean potential temperature model minus observations (Locarnini et al., 2010) difference distributions for the Atlantic Ocean. In MIT, MRI-F, KIEL, CERFACS, CNRM, MRI-A, FSU, and GISS, the difference distributions do not include the high latitude North Atlantic and / or Arctic Oceans, and hence are masked.

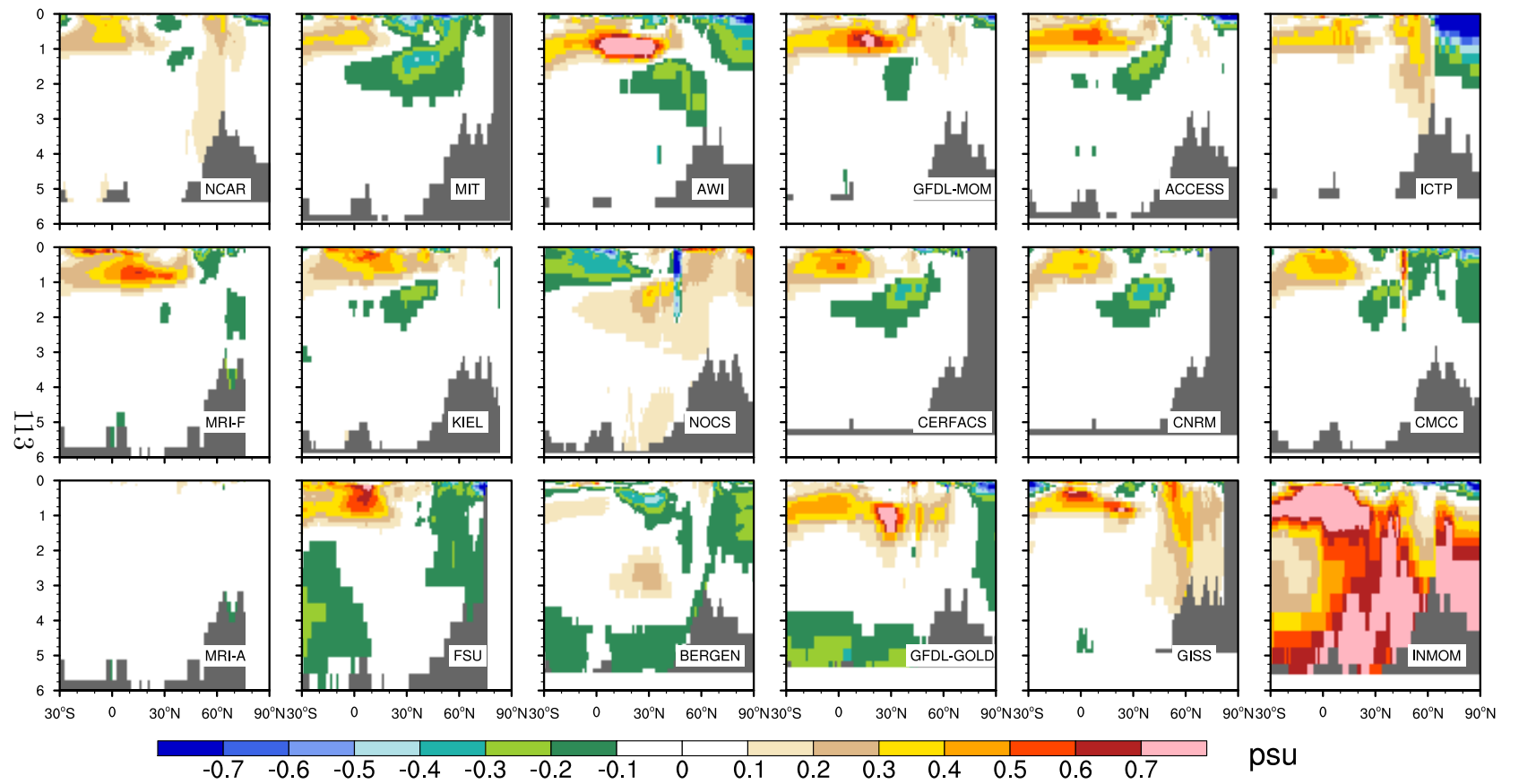


Figure 12: Same as in Fig. 11 except for salinity, and observations are from Antonov et al. (2010)

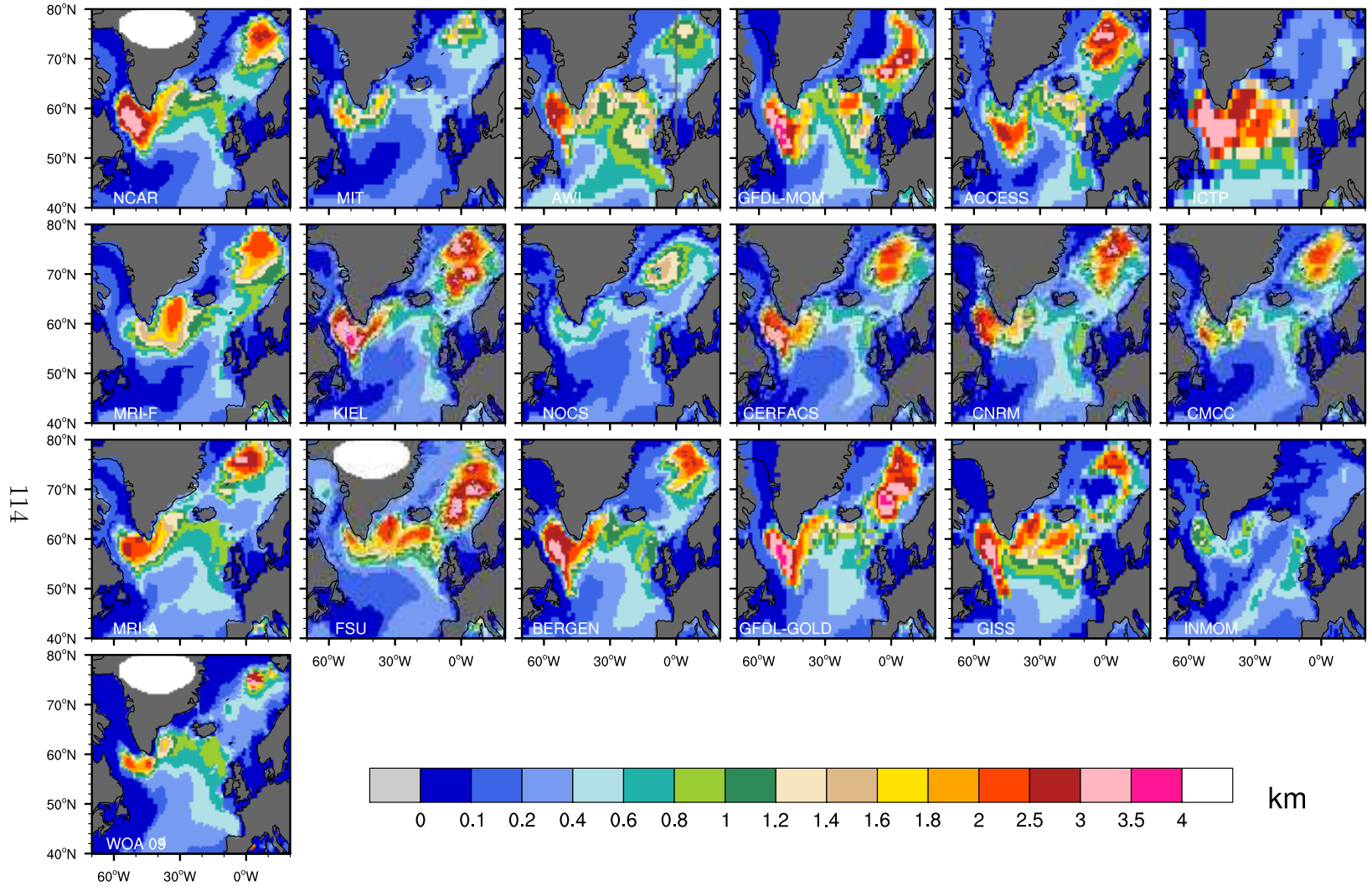


Figure 13: March-mean mixed layer depth (MLD) based on a $\Delta\rho = 0.125 \text{ kg m}^{-3}$ criterion for the northern North Atlantic. The panel to the left of the color bar shows MLD calculated from the WOA09 potential temperature (Locarnini et al., 2010) and salinity (Antonov et al., 2010) data.

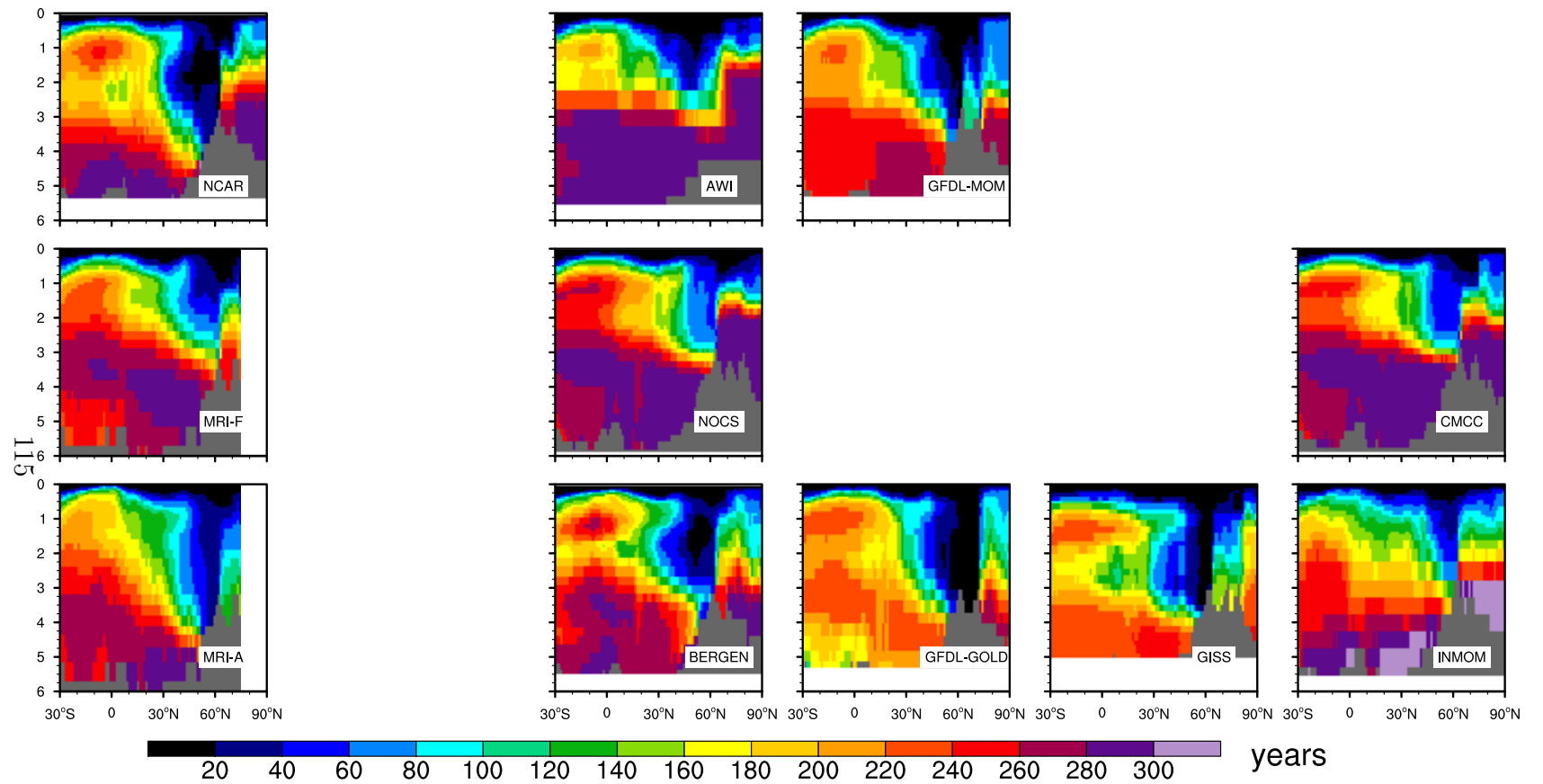


Figure 14: Time-mean and zonal-mean ideal age distributions for the Atlantic Ocean. Ideal age is not available from MIT, ACCESS, ICTP, KIEL, CERFACS, CNRM, and FSU. In MRI-F and MRI-A, the distributions do not include the Arctic Ocean, and hence are masked.

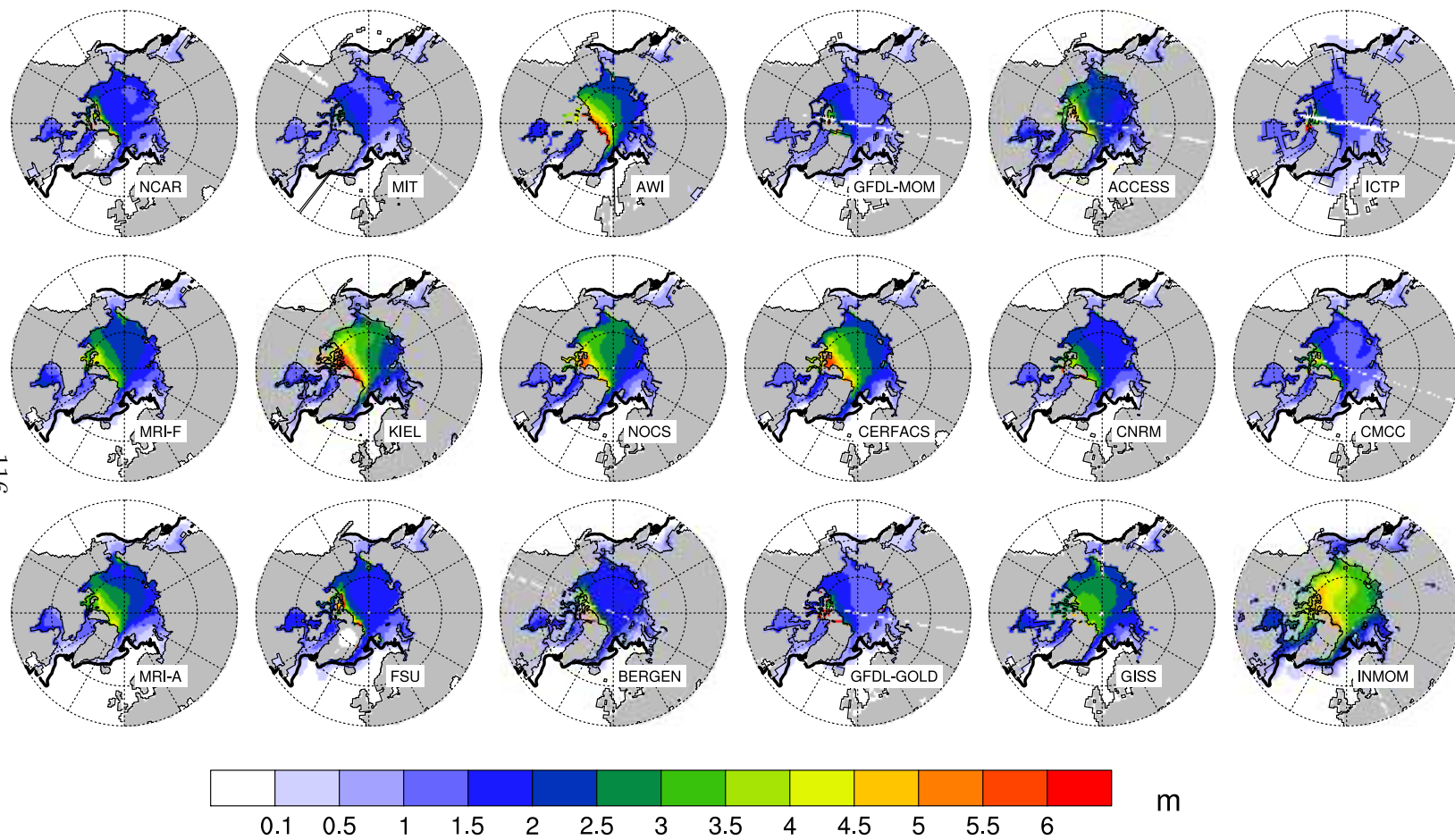


Figure 15: March-mean sea-ice thickness. The black contour line denotes the 15% observed sea-ice extent from the updated NSIDC data (Cavalieri et al., 1996, updated yearly).

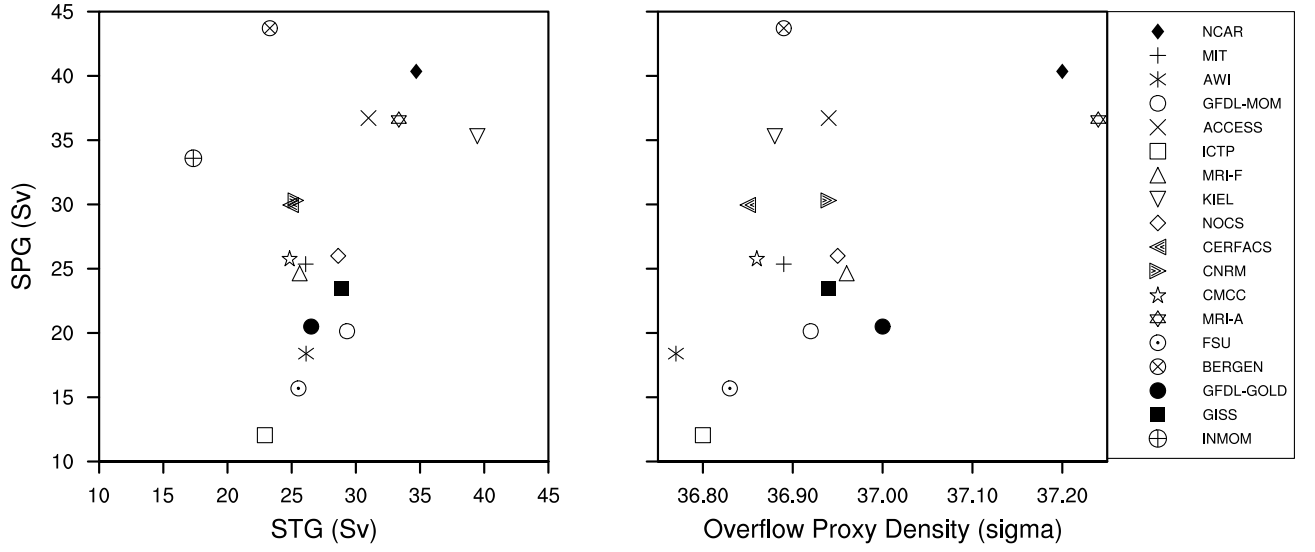


Figure 16: Scatter plots of the North Atlantic subpolar gyre (SPG) maximum transports against the North Atlantic subtropical gyre (STG) maximum transports (left) and against the overflow proxy density (right). All data are time-mean. Overflow proxy density is not available from INMOM. See text for details of how STG and SPG transports are determined. The overflow proxy density is calculated as the time-mean density of the densest outflow (or southward flow) at 60°N as represented by approximately 1 Sv AMOC transport in density (σ_2) space, using Fig. 4.

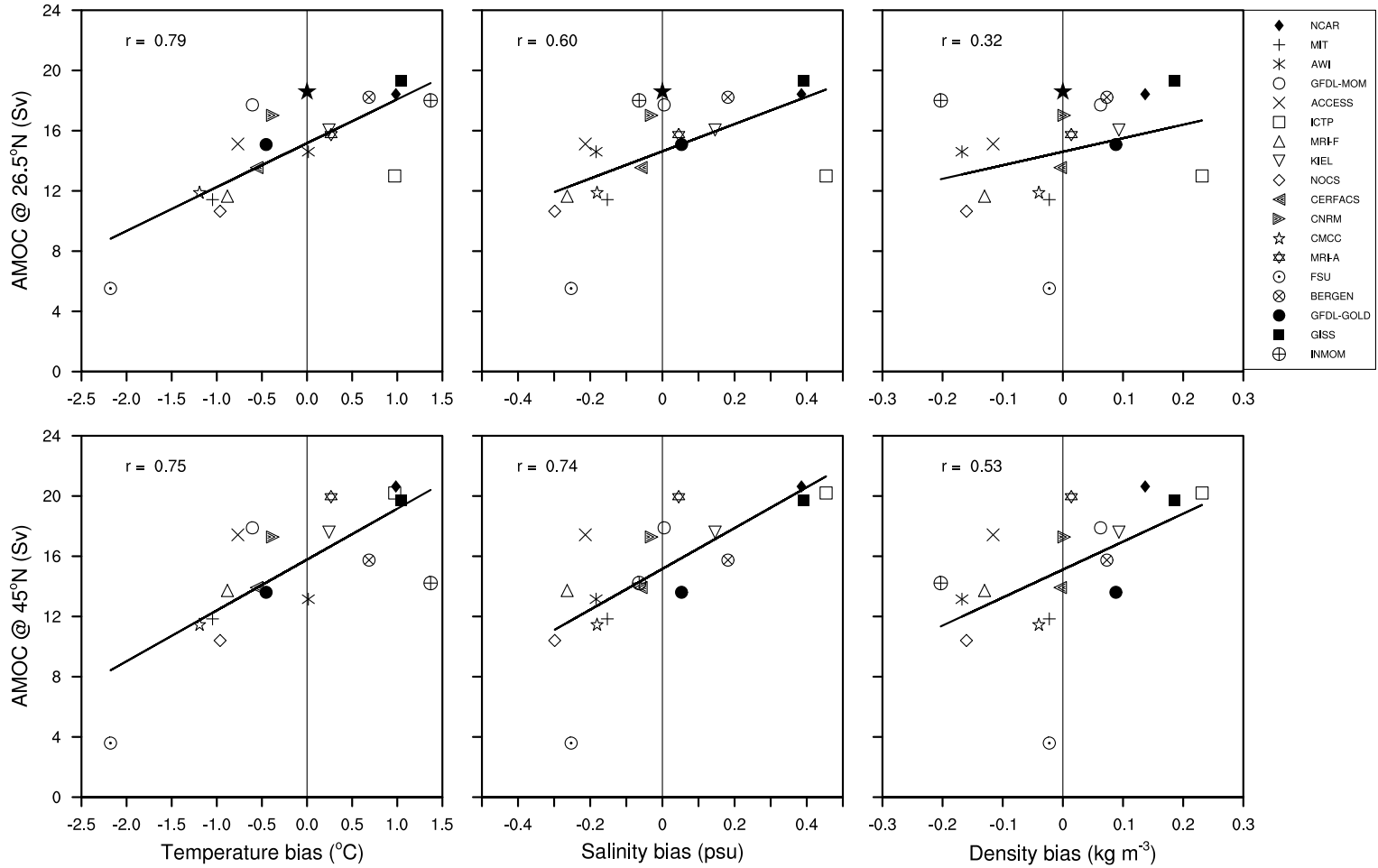


Figure 17: Scatter plots of the Labrador Sea (LS) upper-ocean (0–700 m average) potential temperature (left), salinity (middle), and density (right) biases against the AMOC maximum transports at 26.5°N (top) and 45°N (bottom). Each panel contains the corresponding regression line along with the correlation coefficient. The model data are for the time-mean. The solid stars in the top panels show the 4-year mean RAPID data (April 2004 - March 2008) plotted against no bias. The LS region is bounded by 60°–45°W and 50°–65°N and is indicated in Fig. 8.

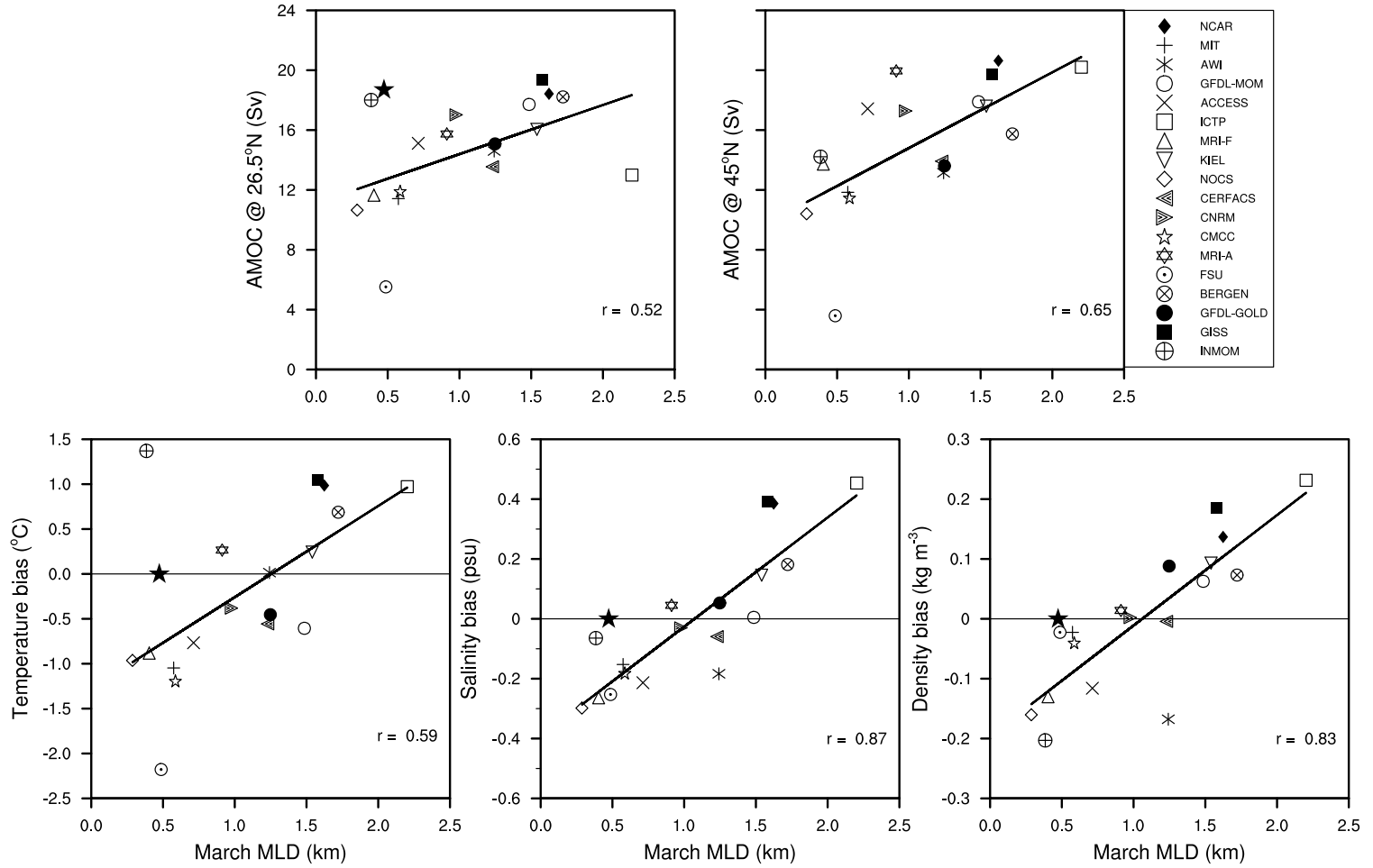


Figure 18: Scatter plots of the LS March-mean MLD against the AMOC maximum transports at 26.5°N and 45°N (top panels) and against the LS upper-ocean (0–700 m average) potential temperature, salinity, and density biases (bottom panels). Each panel contains the corresponding regression line along with the correlation coefficient. Except MLD, the model data are for the time-mean. The solid stars in the top left and bottom panels represent the observationally-based March-mean MLD estimate plotted against the 4-year mean RAPID data (April 2004 - March 2008) and against no bias, respectively.

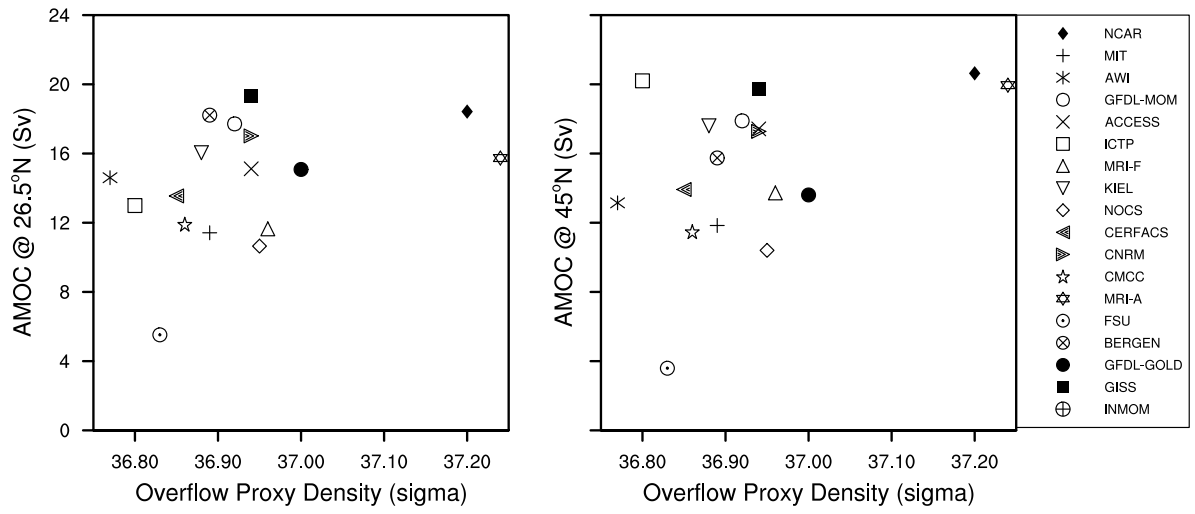


Figure 19: Scatter plots of the AMOC maximum transports at 26.5°N (left) and 45°N (right) against the overflow proxy density described in Fig. 16. All data are time-mean. Overflow proxy density is not available from INMOM.

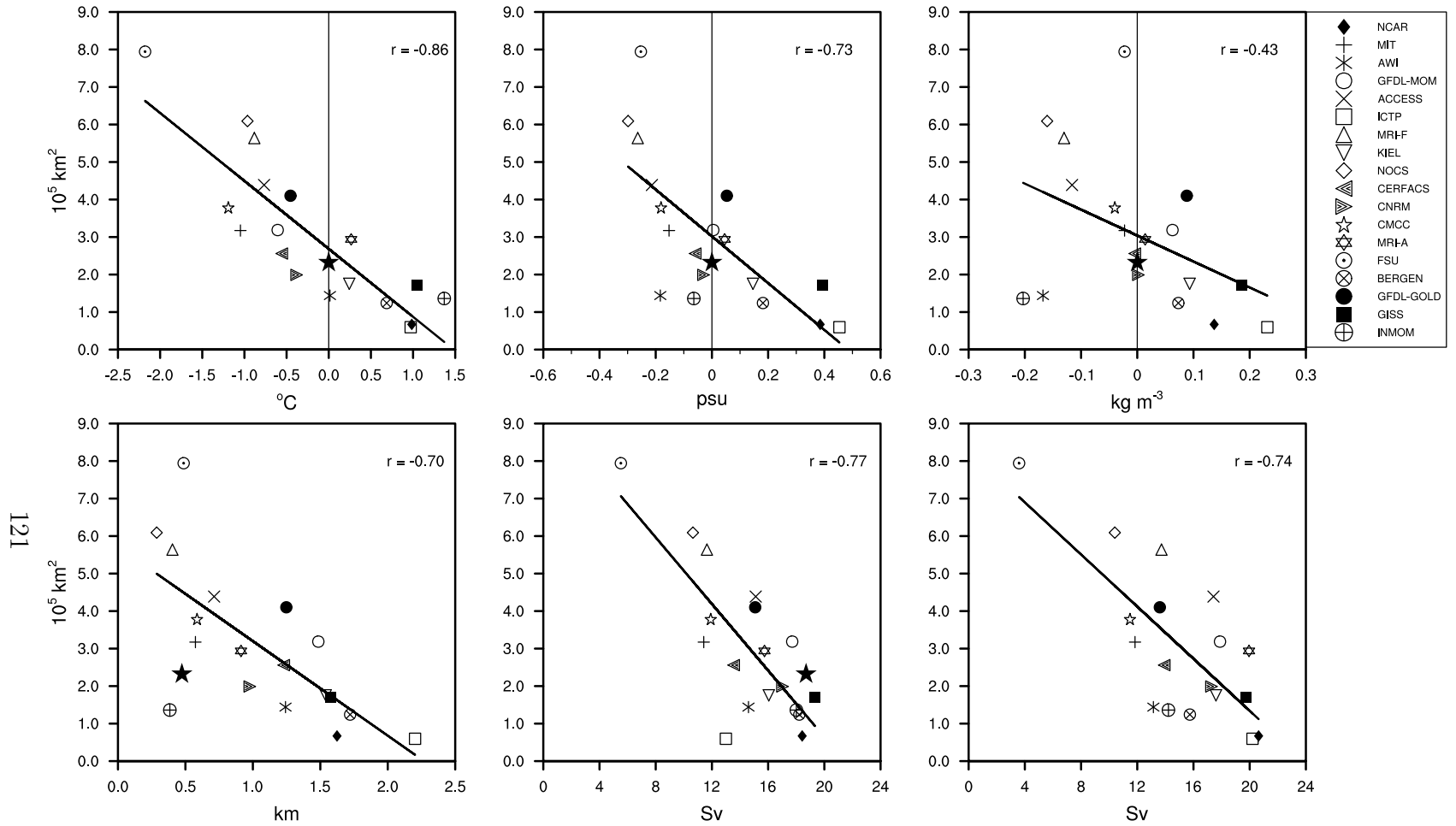


Figure 20: Scatter plots of the LS March-mean sea-ice area against the LS upper-ocean (0–700 m average) potential temperature (top left), salinity (top middle), and density (top right) biases, and against the LS March-mean MLD (bottom left), AMOC maximum transport at 26.5°N (bottom middle), and AMOC maximum transport at 45°N (bottom right). Each panel contains the corresponding regression line along with the correlation coefficient. Except MLD and sea-ice area, the model data are for the time-mean. The solid stars in the top panels show observational March-mean sea-ice area plotted against no bias. The solid stars in the lower left and middle panels are for the observational March-mean sea-ice area plotted against the observationally-based March-mean MLD and the 4-year mean RAPID data (April 2004 - March 2008), respectively.

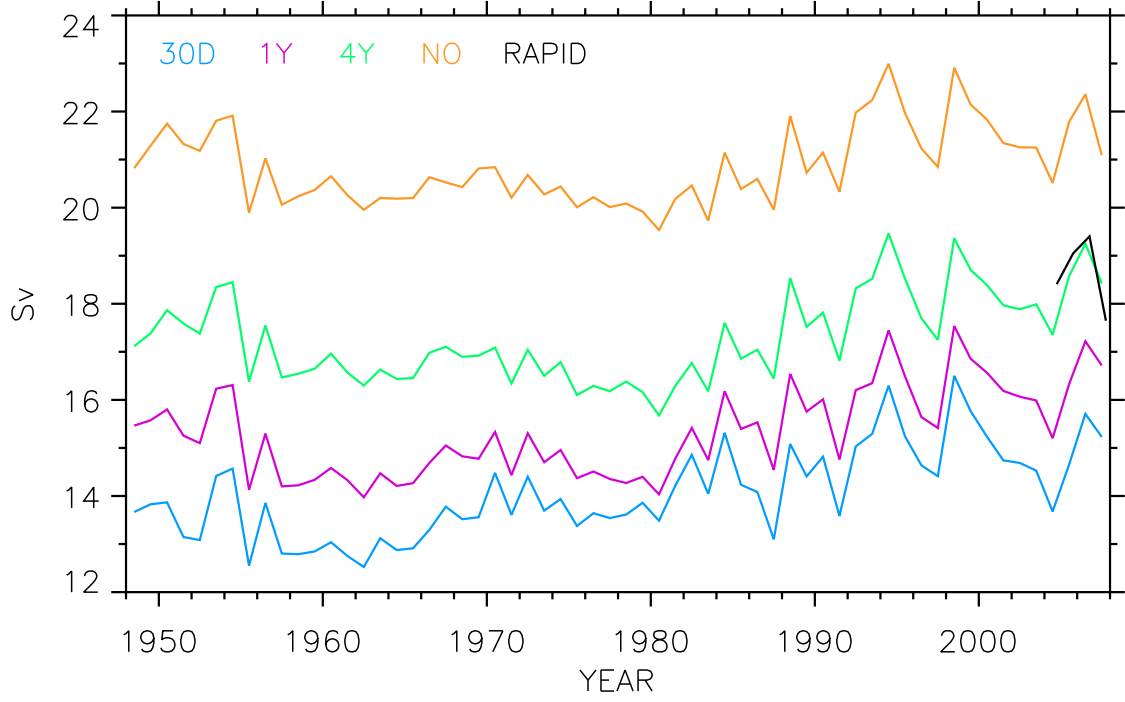


Figure 21: AMOC annual-mean maximum transport time series at 26.5°N obtained with a preliminary version of the NCAR model using different SSS restoring time scales: 30 days (30D); 1 year (1Y); 4 years (4Y); and infinity (NO), i.e., no restoring. The associated length scale is 50 m. RAPID line represents the observational data from Cunningham et al. (2007). The time series are shown for one forcing cycle.

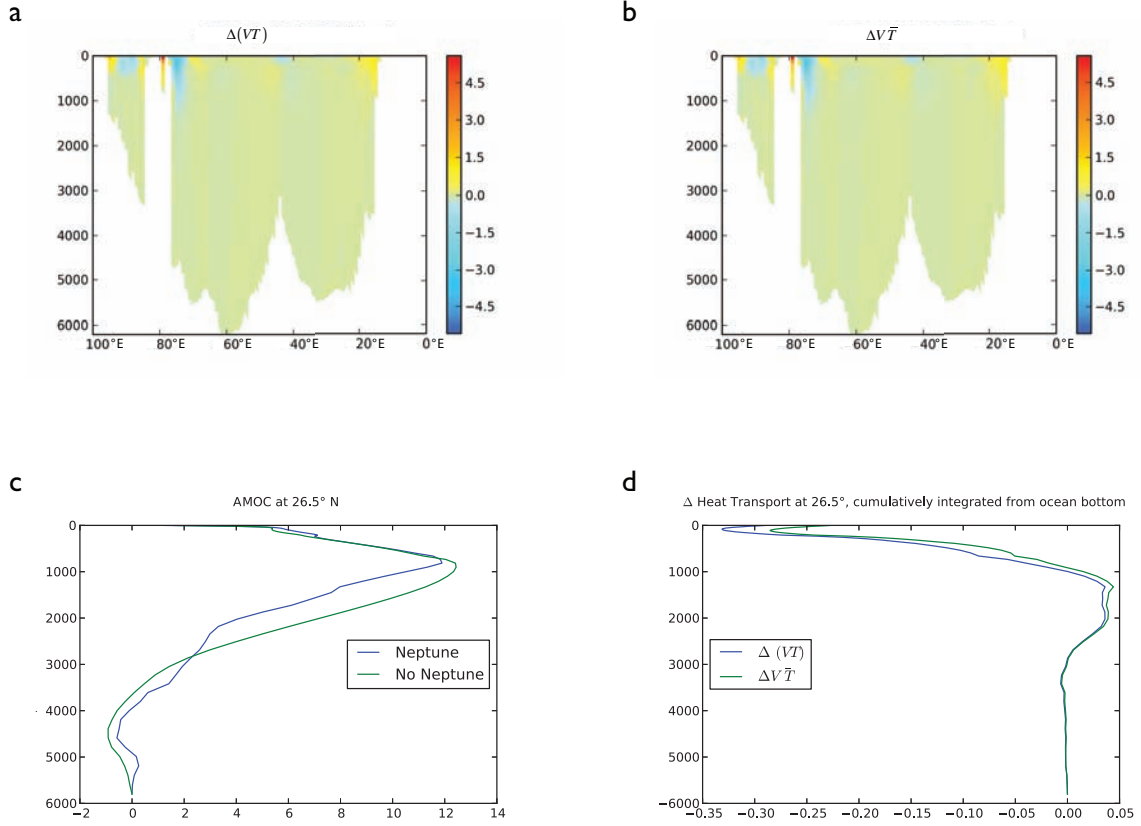


Figure 22: a) Difference (Neptune – standard) in northward annual-mean temperature flux along a quasi-zonal section at 26.5°N in $^\circ\text{C m s}^{-1}$. b) Difference in velocity times mean of the temperatures from the two runs in $^\circ\text{C m s}^{-1}$. c) Zonally integrated meridional flow cumulatively integrated up from the bottom in Sv for the Neptune run (blue line) and standard run (green line). d) Difference in zonally integrated heat transport cumulatively integrated up from the bottom in PW , calculated using the full temperature flux (blue line) and the approximation, using only the difference in velocities (green line). All plots are for year 2007 of the last forcing cycle.

Table 1: Summary of the ocean and sea-ice models in alphabetical order according to the participating group name (first column). The table includes the name of the combined ocean – sea-ice configuration (if any); the ocean model name and its version; the sea-ice model name and its version; vertical coordinate and number of layers / levels in parentheses; orientation of the horizontal grid with respect to the North Pole / Arctic; the number of horizontal grid cells (longitude \times latitude); and the horizontal resolution (longitude \times latitude). In MRI-A and MRI-F, the vertical levels shallower than 32 m follow the surface topography as in sigma-coordinate models. In FESOM, the total number of surface nodes is given under horizontal grid, because it has an unstructured grid. H79 is Hibler (1979) and MK89 is Mellor and Kantha (1989).

Group	Configuration	Ocean model	Sea-ice model	Vertical	Orientation	Horiz. grid	Horiz. res.
ACCESS	ACCESS-OM	MOM 4p1	CICE 4	z^* (50)	tripolar	360×300	nominal 1°
AWI	FESOM			z (46)	displaced	126000	nominal 1°
BERGEN	NorESM-O	MICOM	CICE 4	σ_2 (51+2)	tripolar	360×384	nominal 1°
CERFACS	ORCA1	NEMO 3.2	LIM 2	z (42)	tripolar	360×290	nominal 1°
CMCC	ORCA1	NEMO 3.3	CICE 4	z (46)	tripolar	360×290	nominal 1°
CNRM	ORCA1	NEMO 3.2	Gelato 5	z (42)	tripolar	360×290	nominal 1°
FSU		HYCOM 2.2	CSIM 5	hybrid (32)	displaced	320×384	nominal 1°
GFDL-MOM	ESM2M-ocean-ice	MOM 4p1	SIS	z^* (50)	tripolar	360×200	nominal 1°
GFDL-GOLD	ESM2G-ocean-ice	GOLD	SIS	σ_2 (59+4)	tripolar	360×210	nominal 1°
GISS		GISS Model E2-R		mass (32)	regular	288×180	$1.25^\circ \times 1^\circ$
ICTP		MOM 4p1	SIS	z^* (30)	tripolar	180×96	nominal 2°
INMOM		INMOM		sigma (40)	displaced	360×340	$1^\circ \times 0.5^\circ$
KIEL	ORCA05	NEMO 3.1.1	LIM 2	z (46)	tripolar	722×511	nominal 0.5°
MIT		MITgcm	H79	z (50)	quadripolar	360×292	nominal 1°
MRI-A		MOVE/MRI.COM 3	MK89; CICE	z (50)	tripolar	360×364	$1^\circ \times 0.5^\circ$
MRI-F		MRI.COM 3	MK89; CICE	z (50)	tripolar	360×364	$1^\circ \times 0.5^\circ$
NCAR		POP 2	CICE 4	z (60)	displaced	320×384	nominal 1°
NOCS	ORCA1	NEMO 3.4	LIM 2	z (75)	tripolar	360×290	nominal 1°

Table 2: Summary of the surface freshwater / salt fluxes and salinity restoring choices in alphabetical order according to the participating group name (first column). Salt vs. water column indicates the type of surface fluxes used for hydrological forcing with water and salt denoting real fresh water and virtual salt fluxes, respectively. The sea surface salinity (SSS) restoring time scales are given in days over a 50 m length scale. The NEMO-based models convert salinity restoring to a fresh water flux (denoted as fw in the column). The other groups apply salinity restoring as a salt flux. Region column indicates the region over which the salinity restoring is used. As shown by $|\Delta\text{SSS}| \leq 0.5$, the majority of the models limit the magnitude of mismatch between the model and observed SSS to 0.5 psu. Under sea-ice column shows whether restoring is applied under sea-ice covered areas. Normalize restoring column indicates whether the model subtracts the global mean of restoring fluxes. Finally, normalize P+R-E refers to whether some sort of normalization to P+R-E is applied to reduce drift.

Group	Salt vs. water	Time scale (day)	Region	Under sea-ice	Norm. restoring	Norm. P+R-E
ACCESS	water	150	global ($ \Delta\text{SSS} \leq 0.5$)	yes	yes	yes
AWI	salt	300	global	yes	yes	yes
BERGEN	salt	300	global ($ \Delta\text{SSS} \leq 0.5$)	yes	no	yes
CERFACS	water	300 (fw)	global ($ \Delta\text{SSS} \leq 0.5$)	yes	no	yes
CMCC	water	365 (fw)	global	yes	yes	yes
CNRM	water	300 (fw)	global ($ \Delta\text{SSS} \leq 0.5$)	yes	no	yes
FSU	salt	1460	global ($ \Delta\text{SSS} \leq 0.5$)	yes	yes	yes
GFDL-GOLD	water	50	global ($ \Delta\text{SSS} \leq 0.5$)	yes	yes	yes
GFDL-MOM	water	300	global ($ \Delta\text{SSS} \leq 0.5$)	yes	yes	yes
GISS	water	1250	global	yes ^a	yes	yes
ICTP	water	275	global ($ \Delta\text{SSS} \leq 0.5$)	no	yes	yes
INMOM	salt	365	global	yes	no	no
KIEL	water	1500 (fw)	global ($ \Delta\text{SSS} \leq 0.5$)	no	no	yes
MIT	water	1500	global ^b	yes	yes	yes
MRI-A	water	365	global ^c	yes	yes	yes
MRI-F	water	365	global ^c	yes	yes	yes
NCAR	salt	1450	global	yes	yes	yes
NOCS	water	300 (fw)	global ($ \Delta\text{SSS} \leq 0.5$)	yes	no	yes

^a In GISS, under sea-ice salinity restoring is used only for the grid cells for which the Hadley Center data-set has sea-ice in its 1975-1984 average. The restoring time scale is 42 days.

^b In MIT, model SSS is relaxed to the WOA05 data (PHC3 in the Arctic). These observational data were modified in the North Atlantic by increasing the salinity values by 0.5 psu.

^c In MRI-A and MRI-F, salinity restoring is not used in coastal grid points with sea-ice cover.