

## Controlling High-Latitude Southern Ocean Convection in Climate Models

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1 **Controlling High-Latitude Southern Ocean Convection in Climate Models** 2 3 Achim Stössel<sup>1</sup>), Dirk Notz<sup>2</sup>), F. Alexander Haumann<sup>3</sup>), Helmuth Haak<sup>2</sup>), 4 Johann Jungclaus<sup>2)</sup>, and Uwe Mikolajewicz<sup>2)</sup> 5 <sup>1)</sup> Department of Oceanography, Texas A&M University, College Station, USA 6 7 <sup>2)</sup> Max-Planck-Institut für Meteorologie, Hamburg, Germany 8 <sup>3)</sup> Environmental Physics, Institute of Biogeochemistry and Pollutant Dynamics & Center for 9 Climate Systems Modeling, ETH Zürich, Switzerland 10 11 12 Preprint of: Stössel, A., Notz, D., Haumann, F. A., Haak, H., Jungclaus, J., Mikolajewicz, U. 13 (2015): Controlling high-latitude Southern Ocean convection in climate models, Ocean 14 Modelling, 86, 58-75. doi:10.1016/j.ocemod.2014.11.008. 15 16 17 Abstract 18 19 Earth System Models (ESMs) generally suffer from a poor simulation of the High-Latitude 20 Southern Ocean (HLSO). Here we aim at a better understanding of the shortcomings by 21 investigating the sensitivity of the HLSO to the external freshwater flux and the horizontal 22 resolution in forced and coupled simulations with the Max-Planck-Institute Ocean Model 23 (MPIOM). Forced experiments reveal an immediate reduction of open-ocean convection with 24 additional freshwater input. The latter leads to a remarkably realistic simulation of the distinct 25 water-mass structure in the central Weddell Sea featuring a temperature maximum of +0.5 °C at 26 250 m depth. Similar, but more modest improvements occur over a time span of 40 years after 27 switching from a forced to a coupled simulation with an eddy-resolving version of MPIOM. The 28 switch is accompanied with pronounced changes of the external freshwater flux and the wind 29 field, as well as a more realistic heat flux due to coupling. Similar to the forced freshwater-flux 30 experiments, a heat reservoir develops at depth, which in turn decreases the vertically integrated 31 density of the HLSO and reduces the Antarctic Circumpolar Current to rather realistic values. 32 Coupling with a higher resolution version of the atmosphere model (ECHAM6) yields distinct 33 improvements of the HLSO water-mass structure and sea-ice cover. While the coupled 34 simulations reveal a realistic amount of Antarctic runoff, its distribution appears too concentrated 35 along the coast. Spreading the runoff over a wider region, as suggested in earlier studies to 36 mimic the effect of freshwater transport through icebergs, also leads to noticeable improvements 37 of the HLSO water-mass properties, predominantly along the coast. This suggests that the spread 38 of the runoff improves the representation of Antarctic Bottom Water formation through enhanced 39 near-boundary convection rather than weakened open-ocean convection.

- 40 **1. Introduction**
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43 Earth System Models (ESMs) suffer from a poor representation of the High-Latitude 44 Southern Ocean (HLSO) (e.g., Russell et al., 2006; Sloyan and Kamenkovich, 2007; Weijer et al., 45 2012; Sallée et al., 2013; Heuzé et al., 2013). This was the case for the climate-model 46 simulations carried out for the fourth assessment report of the IPCC (e.g., Lemke et al., 2007; 47 Meehl et al., 2007), and also applies to the simulations of the fifth coupled model 48 intercomparison project (CMIP5) that form the basis for the fifth assessment report IPCC-AR5. 49 The shortcomings manifest themselves e.g. in the fact that CMIP5 ESMs usually simulate 50 excessive interannual variability of the Southern Ocean (SO) sea-ice cover (e.g., Zunz et al., 51 2013; Turner et al., 2013). This is coupled to the representation of winds and the hydrographic 52 properties of the HLSO, which in turn depend on the surface buoyancy fluxes, and thus on the 53 representation of sea ice. In this contribution, we explore the sensitivity of the HLSO in ESMs to 54 ultimately determine the main controlling factors for an improved representation of the water-55 mass properties, of sea ice, and of the strength of the Antarctic Circumpolar Current (ACC). 56 With HLSO we refer to the region south of the ACC, i.e. generally the region south of  $60^{\circ}$ S.

57 Much attention has been given to the question why most ESMs simulate a decreasing SO 58 sea-ice cover in recent decades whereas satellite-derived sea ice shows a slight increase (e.g., 59 Maksym et al., 2012; Parkinson and Cavalieri, 2012). Based on satellite-derived sea-ice drift and 60 atmospheric reanalysis winds, Holland and Kwok (2012) explain the net upward trend in total 61 SO sea-ice area by regional changes in the wind fields. The failure of ESMs to reproduce this 62 upward trend is therefore at least in part linked to their failure to reproduce the observed regional 63 trends in winds (e.g., Haumann et al., 2014b). The model representation of regional winds in 64 general is particularly relevant in Antarctic coastal regions, where the wind fields are strongly 65 affected by orography (e.g., Mathiot et al., 2010; Stössel et al., 2011; Vihma et al., 2011). The 66 coastal wind fields influence the formation of coastal polynyas and thus the formation rate of 67 Antarctic Bottom Water (AABW) through near-boundary convection (e.g., Carmack, 1990). 68 Changes in the volume of AABW affect its northward extension along the deep western 69 boundaries. In the Atlantic it modulates the outflow of North Atlantic Deep Water (NADW) (e.g., 70 Hall et al., 1997; Hogg and Zenk, 1997; Stössel and Kim, 2001; Stössel et al., 2002; Cheon and 71 Stössel, 2009), which in turn influences the long-term properties of Circumpolar Deep Water

72 (CDW) and thus the stratification of the HLSO (e.g., Park and Latif, 2008; Martin et al., 2013;

73 Cheon et al., 2014).

74 Enhanced atmospheric CO<sub>2</sub> causes a poleward shift of the westerlies (e.g., Oke and England, 75 2004; Toggweiler et al., 2006; Fyfe et al., 2007) thus causing a stronger negative wind stress curl 76 over the HLSO. The latter implies a strengthening of the Weddell Sea (WS) and Ross Sea (RS) 77 gyres, a more pronounced doming of isopycnals, and thus a more likely preconditioning for 78 open-ocean convection (e.g., Gordon and Huber, 1990; Cheon et al., 2014). On the other hand, 79 open-ocean convection will decrease in a warmer climate because of changes in the hydrological 80 cycle (de Lavergne et al., 2014; Gordon, 2014). Most directly, warming leads to more 81 precipitation over the SO (e.g., Knutti and Sedláček, 2013). This strengthens the ocean 82 stratification, thus reducing upward ocean heat flux in the HLSO, which in turn will temporarily 83 increase the concentration and/or thickness of sea ice (e.g., Bitz et al., 2005; Zhang, 2007; 84 Gordon et al., 2007; Massonnet et al., 2013). In addition, more snow on sea ice will increase the 85 amount of snow-ice formation, which contributes to sea-ice thickening (e.g., Powell et al., 2005; 86 Maksym and Markus, 2008). Furthermore, while atmospheric warming leads to an overall 87 reduction in ice growth, the associated reduction in oceanic heat flux may reduce ice melt to the 88 point that the overall ice cover turns thicker and more compact in regions of initially strong 89 convection (Zhang, 2007). Kirkman and Bitz (2011) argue that this effect is much stronger than 90 the expected change in precipitation. Another major surface freshwater input in the HLSO is 91 glacial melt water (e.g., Bintanja et al., 2013) though its significance for explaining the recent SO 92 sea-ice increase has been disputed by Swart and Fyfe (2013).

93 Another line of reasoning to explain the increased sea-ice cover in the Southern Ocean 94 relates to internal variability and internally driven long-term trends (e.g., Goosse et al., 2009; 95 Martin et al., 2013). In two multiple-1000-year simulations with the coupled Kiel Climate Model 96 (ECHAM5-NEMO-LIM), Martin et al. (2013) find the mode of open-ocean convection in the 97 WS changing from a stage of deep convection to a stage of no convection and back on time 98 scales of 50 to several 100 years. They explain this by the buildup of a heat reservoir during a no 99 convection phase up to the point when the water column becomes unstable. The transition back 100 to a mode of no convection is explained through gradual surface freshening, the eventual switch 101 being triggered by seasonal sea-ice melt passing a certain threshold value. This study suggests 102 that the trend in sea ice over the last 40 years could simply be due to multi-decadal changes in

open-ocean convection. The CMIP5 analysis of de Lavergne et al. (2014), on the other hand,
suggests that models that are convecting during the preindustrial era all switch to a nonconvecting mode in the future with no point of return (Gordon, 2014).

106 Most ESMs reveal significant biases in HLSO water-mass characteristics (e.g., Orsi and 107 Whitworth, 2005; Russell et al., 2006; Close and Goosse, 2013; Sallée et al., 2013). This issue is 108 to first order linked to the question of how AABW is being formed in individual models, i.e. 109 either predominantly by open-ocean convection, or by near-boundary convection (e.g., Carmack, 110 1990; Marshall and Schott, 1999, Heuzé et al., 2013). According to current understanding and to 111 what has been observed over the past decades, it seems that near-boundary convection including 112 the process of entrainment during downslope flow is the dominant mechanism determining the 113 formation rate and properties of AABW. From that point of view, nature may already have 114 shifted to a permanent mode of no open-ocean convection due to anthropogenic  $CO_2$  emission. 115 Furthermore, models showing excessive convection may not be wrong but rather reflect 116 conditions of the recent past due to different boundary conditions such as a negative Southern 117 Annular Mode (SAM) index (e.g., Thompson and Solomon, 2002; Gordon et al., 2007; Cheon et 118 al., 2014).

119 To better understand the sensitivity of the various interacting processes controlling the 120 distinct HLSO hydrography, we here analyze a set of uncoupled and coupled simulations with 121 different versions of the current state-of-the-science ocean and climate model developed at the 122 Max-Planck-Institute for Meteorology (MPI). This model, as many other CMIP5 models, 123 features excessive open-ocean convection in the HLSO in its standard setup (e.g., Jungclaus et al., 124 2013; Close and Goosse, 2013; Sallée et al., 2013, de Lavergne et al., 2014) compared to HLSO 125 conditions observed over the last 30 years. This deficiency has already been noticed in Griffies et 126 al. (2009), who compared the performance of various sea-ice - ocean GCMs, all using 127 presumably identical forcing fields. A critical variable affecting open-ocean convection in the 128 HLSO is the external freshwater flux, e.g. in form of glacial melt water (e.g., Marsland and 129 Wolff, 2001). We therefore start with investigating this effect in a set of uncoupled simulations. 130 We then analyze the changes that occurred in an existing elaborate eddy-resolving version of the 131 MPI ocean model when switching from uncoupled to coupled simulations, focusing on the short-132 term adjustment of the HLSO water masses over a time span of ~50 years after coupling. Toward 133 the end, this is being compared to long-term results after several 1000 years of coupled

integration that contributed to CMIP5, and the outcome of a sensitivity study on the impact of redistributing glacial meltwater in a coupled simulation. Our strategy is to approach the problem using different model variants (model resolution, initial conditions, forced versus coupled) to eventually determine how to properly balance the dominant processes controlling the HLSO of coupled ESMs.

In the following section 2, we describe our analysis strategy and experimental designs. In section 3, we show results from two sets of numerical experiments, focusing on the spatial distribution of integrated variables such as water-column mean density, mean external freshwater flux, sea-ice thickness and surface heat fluxes. The interaction between these variables and the resulting quality of HLSO water properties is discussed in section 4. In section 5, we summarize the main findings and conclude by investigating the effect of redistributed glacial meltwater in an additional experiment.

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## 148 2. Analysis strategy and model setup

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For our analysis, we use different model setups all including the MPI Ocean Model MPIOM. This model has been used in numerous studies (e.g., Jungclaus et al., 2006; Müller et al., 2010) and forms the sea-ice - ocean component of the MPI-ESM that contributed, for example, to the IPCC-AR5 projections (Jungclaus et al., 2013; Notz et al., 2013). We either use MPIOM in a stand-alone setup forced with atmospheric reanalysis/climatology, or coupled to ECHAM6, which constitutes the state-of-the-science atmosphere model component of MPI-ESM (e.g., Stevens et al, 2013).

157 We also use MPIOM and ECHAM6 at different resolutions. High-resolution simulations 158 revealed substantial improvements in the representation of HLSO water masses and ACC 159 strength (see section 3) upon switching from forced to coupled mode of operation. To understand 160 this behavior, sensitivity experiments were conducted with the low-resolution (1.5 degree) 161 version of MPIOM (MPIOM/GR15, in the following simply called GR15), which served as the 162 ocean component of the low-resolution coupled model MPI-ESM-LR that was used for most 163 CMIP5 experiments (Jungclaus et al., 2013). GR15 has a bipolar grid configuration with the 164 poles over Greenland and Antarctica. Some of our sensitivity experiments were motivated by the 165 study of Griffies et al. (2009), where GR15 was one of 4 models (of a total of 7) that delivered

166 excessively deep mixed layers south of the ACC (their Fig.15 top right) and a substantial over-167 estimation of the ACC strength (185 Sv; estimates derived from measurements amount to 168 135±20 Sv (e.g., Whitworth and Peterson, 1985; Cunningham et al, 2003). In addition to the 169 effect of salinity restoring (not shown), we investigated the impact of additional freshwater input, 170 in particular following the hypothesis that the external freshwater supply provides a dominant 171 control over convection and thus the overall water-mass properties of the HLSO. The reference 172 simulation without additional freshwater flux is termed "REF". Besides precipitation, the 173 freshwater input into the HLSO is to a large extent determined by runoff from the Antarctic 174 continent in form of glacial meltwater and icebergs (e.g., Gladstone et al., 2001; Beckmann and 175 Goosse, 2003; Silva et al., 2006; Martin and Adcroft, 2010; Jongma et al., 2009; Tournadre et al., 176 2012; Bintanja et al., 2013; Weber et al., 2014). There are large uncertainties in the estimates of 177 the rate and spread of glacial melt water from Antarctica. Gladstone et al. (2001) reported 178 regional values south of  $65^{\circ}S$  of up to 0.5 m/year. Deriving iceberg trajectories by means of 179 satellite remote sensing, Tournadre et al. (2012) estimate regional freshwater fluxes due to 180 melting icebergs north of 65°S up to 0.3 m/year. To investigate the possible impact of an overall 181 larger freshwater flux, we followed Marsland and Wolff (2001) by simply adding a fixed rate of 182 freshwater input south of 60°S, estimating an upper bound of the possible response by using an 183 unrealistically high value of 1 m/year (experiment "FW").

184 The atmospheric forcing field used for the GR15 experiments are based on the OMIP 185 climatology (Röske, 2006). These "GR15-OMIP" experiments start from rest with climatological 186 ocean properties according to NODC\_WOA98 (http://www.esrl.noaa.gov/psd/) and were run for 187 50 years (Table 1). This is clearly not enough time for the global overturning circulation to adjust, 188 but sufficient for our main aim to investigate the initial adjustment of the HLSO density structure 189 and the tendency of the strength of the ACC without the long-term impact of possible changes in 190 the outflow of NADW that would ultimately affect the abundance and property of CDW. The 191 two forced experiments are subject to salinity restoring where there is no ice. The associated time 192 constant can be converted into a "piston velocity" [m/s] which translates into a freshwater flux q 193 = -  $v_p (S_{obs} - S_{mod})/S_{mod}$ , where  $v_p$  is the piston velocity,  $S_{obs}$  is the observed salinity, and  $S_{mod}$  is the modeled salinity (Griffies et al., 2009). Both forced GR15 experiments use  $v_p = 3.3 \cdot 10^{-7}$  m/s, 194 195 which corresponds to a restoring time scale of 1.2 years over the upper-layer depth of 12.5 m.

196 The high-resolution experiments we examine are based on a tripolar (TP) grid MPIOM 197 version at 0.1 degree or 6 minutes (6M) horizontal resolution, MPIOM/TP6M (von Storch et al., 198 2012), in the following simply called TP6M. This high-resolution model setup features a 199 substantially higher resolution of the bathymetry than GR15 and than the MPIOM version used 200 in the CMIP5 intermediate-resolution setup MPI-ESM-MR (TP04, i.e. tripolar 0.4 degree 201 horizontal resolution), in particular since all MPIOM versions feature partial grid cells (e.g., 202 Maier-Reimer et al., 1993; Adcroft et al., 1997). This leads to a more detailed representation of 203 the accumulation and overflow of dense shelf water (e.g., Orsi et al., 1999; Heuzé et al., 2013). 204 The TP6M experiments we examine here constitute a sequence of simulations that started with 205 25 years of OMIP-climatology-forced MPIOM/TP6ML80 (from rest with NODC\_WOA98 206 ocean properties), where L80 stands for 80 vertical levels. This simulation continued for another 207 60 years with atmospheric forcing based on real-time 6-hourly NCEP reanalyses (J.v.Storch et al., 208 2012). The end of that forced simulation was the starting point for 2 coupled simulations, namely 209 MPIOM/TP6ML40-ECHAM6/T255 (in the following called TP6M-T255) and 210 MPIOM/TP6ML40-ECHAM6/T63 (in the following called TP6M-T63), where T255 stands for 211 spectral triangular truncation at wave number 255, and T63 correspondingly at wave number 63 212 (e.g., Washington and Parkinson, 2005).

Finally, the results of the forced GR15 and the coupled TP6M experiments are compared with those of the preindustrial control (piCon) simulations of MPI-ESM-LR (in MPI model name terminology MPIOM/GR15L40-ECHAM6/T63, and for our comparison simply GR15-T63piCon). The piCon integrations were run for several thousand years in coupled mode with constant radiative forcing at preindustrial level. In these simulations, the global ocean overturning circulation and deep-ocean properties have equilibrated, and the HLSO properties will include the long-term effect of changes in NADW outflow and properties.

Originating from GR15-T63-piCon, a coupled runoff experiment (GR15-T63-RO) was added to investigate the impact of spreading the Antarctic runoff (or glacial meltwater) over the entire ocean region south of 60°S. This is a crude representation of the spreading of melt water through icebergs. A more adequate implementation would be that of e.g. Martin and Adcroft (2010) where interactive icebergs are simulated as Lagrangian particles in an Eulerian framework.

226 For experiment identification we omitted "MPIOM" for forced experiments, and "MPIOM-227 ECHAM6" for coupled experiments, i.e. just indicated the MPIOM horizontal resolution and 228 grid (e.g., GR15, TP6M), followed by the forcing (e.g., OMIP, NCEP) for the uncoupled 229 experiments, and the resolution of ECHAM6 for the coupled experiments (e.g., T255, T63). 230 Supplemental model experiments are flagged with REF (REFerence), FW (FreshWater), and RO 231 (RunOff). Separated by a dot, this is augmented by the simulation year being investigated (e.g., 232 year 40), or the actual year of real-time forcing (e.g., year 2010 of NCEP). All discussed 233 experiments are listed in the left column of Table 1.

234 We mostly show results from the final simulation year of the sensitivity experiments. These 235 are directly compared to the archived results of the TP6M experiments 40 to 90 years into 236 coupled integration (i.e., as long as these expensive experiments have been run). Coupled 237 simulations are generally prone to higher interannual variability than forced simulations, in 238 particular for variables such as sea-ice concentration. Focusing on winter values we found that 5-239 year August means did not differ much from single-year August means, in particular not to the 240 point that they would affect our main conclusions. For sea ice we nevertheless show 5-year 241 August means, i.e. of the last 5 years of the respective sensitivity experiments or the respective 242 stage of the TP6M experiments. The shown wind stress was treated similarly, but reflects 6-year 243 means over entire years. Furthermore, the annual-mean ACC strength is shown as time series.

244 To analyze the quality of the simulations, we focus on a limited number of characteristic SO 245 quantities. These include 2-d horizontal maps of vertical-mean density below 200 m, sea-ice 246 thickness, external freshwater flux (defined as precipitation minus evaporation plus continental 247 runoff of any sort, i.e. not including freshwater flux from thermodynamic sea-ice growth and 248 melt), surface heat flux, and near-surface air temperature. 3-d fields that we examined include 249 potential temperature, salinity, and potential density. An important indicator for model 250 performance of the HLSO is the meridional water-mass distribution across the WS. A reliable 251 source for its verification is WOCE repeat section A23 (e.g., Orsi and Whitworth, 2005). Since 252 the meridional density gradient across the ACC is directly linked to its strength, this is also being 253 monitored for each experiment. Other quantities are the meridional distribution of zonal wind 254 stress and the wind-stress curl.

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- 258 **3. Results**
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260 Fig.1 shows meridional sections of potential temperature, salinity and potential density along 261 30°W from the WOCE climatology, providing a typical picture of the HLSO water-mass 262 properties, in this case through the WS gyre. The upper 50 to 100 m are characterized by 263 temperatures close to the freezing point and relatively low salinity (note that measurements are 264 generally summer biased). Deeper down is the core of the relatively warm (up to +0.5 °C) and 265 saline (up to 34.69 g/kg) CDW at about 300 m depth extending all the way to the Antarctic 266 continental margin where this core deepens to about 800 m. The density structure is mainly 267 determined by the salinity distribution and the dynamics of the WS gyre, leading to a doming of 268 isopycnals with a potential density maximum (referenced to 0 dbar) up to 1027.86 kg/m<sup>3</sup>.

269 The corresponding sections from August of TP6M-NCEP.2010 look distinctly different (Fig. 270 2). Irrespective of possible seasonal and interannual variability in the upper 200 m, the pattern 271 shown in the figure is typical for forced MPIOM simulations (e.g., Griffies et al., 2009), 272 illustrating the problem of excessive open-ocean convection. There are clearly major 273 discrepancies to the measurement-based climatology. While fronts across the ACC are readily 274 discernable and constitute a realistic feature, the central WS gyre is completely mixed with no 275 indication of CDW whatsoever. Furthermore, the temperature is at the surface freezing point 276 throughout the water column. Salinity is vertically homogeneous at 34.7 g/kg. As a result of the 277 excessively cold temperature throughout the water column, the density (>  $1029.0 \text{ kg/m}^3$ ) is 278 overestimated across much of the WS gyre. Note that the focus here is the water column below 279 200 m, i.e. below the part being affected by seasonality.

280 Table 1 lists all experiments discussed in this study together with a number of critical 281 quantities. In TP6M-NCEP.2010 the strength of the ACC reaches up to 230 Sv. We note a 282 remarkable improvement of the ACC strength upon coupling (TP6M-NCEP.2010 vs. TP6M-283 T255.40). A likely candidate for this improvement is the HLSO density that eventually 284 determines the meridional density gradient across the ACC, and thus its strength (e.g. Pierce et 285 al., 1995; Martin et al., 2013). To investigate this relationship, we calculated the 3d-spatially 286 averaged August-mean density below 200 m south of 70°S and south of 60°S, respectively, which 287 is given in Table 1 along with the annual-mean ACC strength. We neglect the upper 200 m so 288 that changes in density are just due to changes in dense-water accumulation rather than due to the

289 spreading of surface freshwater. For selected experiments, the ACC strength is also shown as 290 time series over the respective integration periods (Fig.3). Within the respective sets of 291 experiments (separated by bold lines in Table 1) there is generally a close relationship between 292 the HLSO density and the ACC strength, i.e. the denser the water south of the ACC, the stronger 293 the ACC. Such relationship has also been noticed by Meijers et al. (2012) and Sallée et al. (2013), 294 and is most pronounced for density south of 70°S. Observed estimates suggest a mean potential 295 density value below 200 m south of 60°S of 1027.81 kg/m<sup>3</sup>, which compares favorably with the 296 simulated HLSO density of the experiments with the lowest and most realistic strength of the 297 ACC. On the other hand, the corresponding observed density south of  $70^{\circ}$ S being the same as 298 that south of  $60^{\circ}$ S does not support it being strongly related to the ACC strength. In fact, this lack 299 in meridional difference also holds for all forced experiments, suggesting that the less dense 300 conditions south of  $70^{\circ}$ S in the coupled experiments have to do with the treatment of the 301 Antarctic runoff (see discussion). The coupled GR15-T63 set of experiments will be discussed in 302 section 4 and 5.

303 To investigate the distribution of density in more detail, we examine the horizontal 304 distribution of vertical- and August-mean density below 200 m (Fig.4). For a convenient 305 comparison, the coarse-grid GR15 results and the fine-grid TP6M results have been interpolated 306 onto a latitude-longitude grid that corresponds to the intermediate-resolution TP04 grid. GR15-307 OMIP-REF.50 yields high density in the central WS and RS gyres, as well as in the high-latitude 308 embayments. This situation changes drastically when adding freshwater south of 60°S (GR15-309 OMIP-FW.50). This measure leads to a substantial reduction of density all along the Antarctic 310 coastline, in particular including the high-latitude embayments. At the same time, this 311 experiment features a reduction of the ACC from 195 Sv to 142 Sv within 50 years (Fig.3).

Moving on to the eddy-resolving TP6M experiments, the density distribution of TP6M-NCEP.2010 resembles that of GR15-OMIP-REF.50, but the former yields overall noticeably denser water. Accordingly, the ACC strength of TP6M-NCEP.2010 is at 230 Sv even more overestimated than that of GR15-OMIP-REF.50 (Table 1). After 40 years of coupled integration the HLSO density has diminished substantially, in particular in the high-latitude embayments, similar to GR15-OMIP-FW.50. The density distribution of TP6M-T63.40 looks rather similar (not shown), as does the total density (Table 1). The ACC is consistently reduced to acceptable values (Table1; Fig.3). Except for the high-latitude embayments, comparison with the WOCE
data indicates that GR15-OMIP-FW.50 yields the most realistic result.

321 The multi-year August-mean sea-ice thickness is shown in Fig.5. It provides a good 322 indication of open-ocean convection and the underlying water-mass structure due to its direct 323 response to the subsurface heat flux. In terms of sea-ice thickness, a measure of good 324 performance is the relative thinness (about 0.5 m) of the majority of the wintertime ice pack (e.g., 325 Worby et al., 2008; Zwally et al., 2008; Stössel, 2008). GR15-OMIP-FW.50 shows a clear 326 increase of ice thickness versus GR15-OMIP-REF.50, which is due to enhanced stratification 327 and reduced ocean heat flux. Upon coupling of TP6M, there is a similar tendency toward a 328 thicker and more compact winter ice cover. On the other hand, there are regions in the eastern 329 WS that are indicative of open-ocean polynyas in TP6M-T63.40 that do not emerge in TP6M-330 T255.40. Furthermore, the winter sea-ice cover of TP6M-T255 is more extensive than that of 331 TP6M-T63. Noting that the coupling frequency in these two experiments is the same (hourly), 332 these differences are only explicable through effects of the different spatial resolution of the 333 atmosphere model. Except for the lacking RS polynya, GR15-OMIP-FW.50 yields the most 334 realistic winter ice thickness, followed by TP6M-T255, which, however, suffers from an 335 excessive extent in the Atlantic sector and too thick ice along East Antarctica.

336 To follow up on the hypothesis that the improvement of the ACC strength upon coupling is 337 mainly due to changes in the HLSO freshwater supply, Table 1 contains the spatially averaged 338 annual-mean external freshwater flux broken down in precipitation minus evaporation and runoff 339 (and implied freshwater flux from salinity restoring in the forced experiments), without 340 considering contributions from sea-ice melt or growth. Except for TP6M-T255.0, which has not 341 experienced any noticeable adjustment, there is a tendency for HLSO density below 200 m to 342 decrease with an increase in external freshwater input mainly when both are restricted to south of 343 70°S, again suggesting that this may be an artifact of the treatment of the Antarctic runoff in the 344 coupled models.

Fig.6 shows a considerable change in external freshwater flux between coupled and uncoupled experiments, and among the uncoupled experiments themselves (grey regions indicate negative freshwater flux). Switching from the uncoupled to the coupled TP6M simulations, the freshwater input into the SO generally diminishes except right along the Antarctic coastline and most of the WS, where the freshwater input actually increases (this is also reflected in Table 1 for freshwater input south of  $70^{\circ}$ S). Even though the overall freshwater input south of  $60^{\circ}$ S decreased, the HLSO turns less dense in the course of the 40 years of coupled integration, suggesting that a large amount of the freshwater south of  $70^{\circ}$ S is being advected northward where it potentially affects the stratification to the point that less dense water is being produced.

354 The forced simulations GR15-OMIP-REF.50 and TP6M-NCEP.2010 show extraction of 355 freshwater (grey regions in Fig.6), predominantly because of salinity relaxation. GR15-OMIP-356 FW.50 shows large positive values south of 60°S, as expected. Note that the salinity restoring 357 acts against the freshwater anomaly in the amount of about 0.5 m/yr (Table 1). The coupled 358 simulations, on the other hand, reveal a strong freshwater input along the coast of Antarctica. 359 Initially, this leads to thinner ice, which recovers during the course of the coupled integration to 360 thicker ice than in the forced simulation (Fig.5), in particular in TP6M-T255. Further offshore, 361 there is apparently more freshwater input into the SO in TP6M-NCEP.2010 than in the coupled 362 TP6M simulations, except for the southern WS and RS. TP6M-T63 shows a coarser grid pattern 363 because of the coarser AGCM resolution. Apparently the coarser grid leads to larger areas 364 around Antarctica being exposed to enhanced freshwater flux than in TP6M-T255, though this is 365 not reflected in the spatially averaged freshwater flux (Table 1).

366 The August-mean surface heat flux (not shown) is highly dependent on the ice cover, with 367 the pattern of strong heat loss in the HLSO following that of low ice thickness and concentration. 368 There is a considerable decrease of heat loss in GR15-OMIP-FW.50 versus GR15-OMIP-369 REF.50 as well as with a switch from uncoupled to coupled mode. The latter is to a large extent 370 due to changes in the turbulent heat fluxes. These depend, besides wind speed, on the 371 temperature and humidity differences between air and sea surface. In the coupled simulation, the 372 air temperature and humidity are able to adjust to the surface conditions, which reduces the 373 vertical gradient and hence the cooling due to turbulent heat fluxes. In the uncoupled case, the air 374 temperature and humidity will remain at their specified values irrespective of the simulated 375 surface conditions. This leads in winter to a situation of strong cooling in areas that are ice free 376 or covered by thin ice. This difference can be seen in the August-mean 2-m air temperature 377 (Fig.7). The region of near-surface air temperature colder than  $-30^{\circ}$ C of the NCEP data is 378 obviously substantially larger than that of TP6M-T255.0, in particular in the RS, the WS, and 379 along East Antarctica. The reduced cooling in the coupled simulations is a main contributor to 380 reducing thermodynamic ice growth, and thus convection and dense-water formation. Augustmean sea-surface salinity (Fig.8) responds not only directly to the external freshwater flux, but also to brine and freshwater release due to freezing and melting. There is a distinct reduction of salinity in GR15-OMIP-FW.50. The reduction in surface salinity is even more pronounced in the sequence TP6M-NCEP.2010, TP6M-T255.40. While TP6M-T63 maintains equally low salinity near the coast of Antarctica, it yields higher salinity in the open ocean compared to TP6M-T255. The lower salinity along Antarctica in both TP6M-T255.40 and TP6M-T63.40 can be attributed to the larger runoff (Fig.6), with salinity values similar to GR15-OMIP-FW.50.

388 Since standard forced MPIOM simulations are characterized by excessive open-ocean 389 convection in the HLSO, it is revealing to investigate the temperature, salinity, and density 390 profiles in the central WS (Fig.9 top row), in the wider WS (Fig.9 middle row), and in the 391 circumpolar HLSO between 60°S and 70°S (Fig.9 bottom row). These will first be analyzed for 392 the TP6M simulations (colored solid lines except purple). Considering that the sequence of 393 TP6M experiments started from an ocean climatology that reflects central WS profiles similar to 394 the solid black lines in Fig.9, it is remarkable that the temperature stratification in the WS 395 disappears completely during the first 25 years of the OMIP-forced phase, at which point the 396 temperature profile is totally homogeneous (not shown). Furthermore, the entire water column 397 attains values close to the surface freezing point. After 60 years of 6-hourly real-time NCEP forcing a slight stratification develops, but with a maximum temperature of only -1.7 °C at 200 m 398 399 depth (not shown). Immediately upon coupling (Fig.9 top left, solid red line), a more pronounced 400 wintertime maximum of -1.4 °C emerges at 250 m depth, which after 40 years of coupled 401 simulation develops into a much more pronounced maximum of -0.2 °C at 300 m (solid orange 402 line). By then the temperature profile has gained a shape much more similar to what is being 403 observed, except that the maximum is still 0.7 °C shy of the observed value. After 40 years of 404 coupled simulation with the coarser AGCM (TP6M-T63.40), the shape of the temperature profile 405 at that location is less realistic in that the maximum reaches only -0.8 °C, and is spread out from 406 400 m to 1500 m (solid dark blue line) rather than being more confined around a water depth of 407 300 m. After another 50 years of integration (TP6M-T63.90), the deeper layers are warming 408 toward more realistic values (solid light blue line), which would presumably also happen in a 409 corresponding continuation of TP6M-T255. However, there is no improvement of TP6M-T63.90 410 versus TP6M-T63.40 above 1700 m. The wider WS (Fig.9 middle left) shows similar behavior, 411 while in the circumpolar belt (Fig.9 bottom left) TP6M-T63 shows a deterioration above 1800 m

upon further integration. The inversion in the top 100 m of the WOCE profile stems from the fact
that observed data is summer biased. The simulations also show a shallow stratification in
summer (not shown).

415 As for salinity in the central WS (Fig.9 top middle), the observed profile (solid black line) 416 shows a summer bias in that its surface salinity is 0.2 g/kg lower than what would be expected in 417 winter. This has been judged from the monthly Polar Hydrographic Climatology (PHC, Ermold 418 and Steele, psc.apl.washington.edu/Climatology.html) ocean data, which is available for that 419 location up to a depth of 1400 m. Otherwise, the PHC profile is similar to the WOCE profile, in 420 particular with respect to displaying a salinity maximum close to 34.69 g/kg at 300 m depth. 421 Together with the temperature maximum at that depth, this is indicative of CDW. While the 422 OMIP-forced spin-up run ends with a realistic salinity at depth (not shown), this high salinity 423 value extends all the way to the surface, again not showing any indication of stratification. 424 During the 60 years of the NCEP-forced simulation, a weak salinity gradient builds up 425 throughout the water column. By the end of that simulation, the salinity at depth has reached a 426 value that is overestimated by 0.05 g/kg. One year into the coupled simulation (TP6M-T255.0), a 427 weak halocline emerges (solid red line) that develops into a pronounced halocline after 40 years 428 of coupled simulation (solid orange line). Furthermore, a maximum of about 34.69 g/kg develops, 429 similar to the observed profiles, though substantially deeper and less confined, while a 430 substantially higher salinity than observed establishes at depth. TP6M-T63.40 features less 431 realistic results (solid dark blue line) as there is a much less pronounced halocline. TP6M-T63.90 432 shows a somewhat improved salinity profile (solid light blue line), though not to the extent of 433 TP6M-T255.40. For the wider WS (Fig.9 middle panel), the finding is similar. On the larger 434 scale (Fig.9 bottom middle) it also seems that TP6M-T255.40 delivers the most realistic profile 435 shape.

The stratification is ultimately indicated by the vertical density distribution (Fig.9 right column). The reference pressure for the density profiles is 0 dbar. At temperatures around the freezing point the summer bias of 0.2 g/kg in salinity translates to about 0.16 kg/m<sup>3</sup> in density (e.g., Talley et al., 2011), i.e. the observed surface density (solid black line) in the central WS should be around 1027.7 kg/m<sup>3</sup> in winter. Focusing first on the central WS, the NCEP-forced simulation yields a realistic density at depth, but there is no indication of any stratification. 60 years into the NCEP-forced integration, stratification develops, but only by deep water becoming too dense (not shown). Immediately upon coupling (TP6M-T255.0), the surface density diminishes (Fig.9 top right, solid red line) in accord with salinity. 40 years into the coupled integration, density at depth returns to more realistic values (solid orange line). The upper layers experience a noticeable stratification in TP6M-T255.40, whereas that of TP6M-T63.40 is much less pronounced (solid dark blue line), and does only improve at depth after another 50 years of integration (solid light blue line). The tendencies are similar in the wider WS (Fig.9 middle right), and apply in this case also to the circumpolar belt (Fig.9 bottom right).

450 The dashed lines in Fig.9 (dark green versus light green) illustrate the impact of the 451 additional freshwater flux in the forced GR15 experiments. In the central WS (Fig.9 top row), 452 GR15-OMIP-FW.50 improves the density profile considerably, in particular in the upper 500 m. 453 The salinity profile of GR15-OMIP-FW.50 is the most realistic of all experiments. The only 454 noticeable difference to the observed profile is the location of the salinity maximum, which is 455 located at 800 m depth, as opposed to 300 m in the observation. The ultimate improvement is 456 found in the temperature profile. Similar improvements are apparent in the wider WS (Fig.9 457 middle row) and in the circumpolar belt (Fig.9 bottom row).

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## 460 **4. Discussion**

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462 Having so far mostly concentrated on individual variables, we will now move on to discuss 463 in more detail the inter-relationship between the variables. We start with an analysis of the 464 factors that influence sea-ice thickness. We find that the deep warm water reservoir and the 465 amount of vertical mixing readily affect the sea-ice thickness, as illustrated in Fig.5. At the end 466 of the forced TP6M experiment and at the beginning of the coupled experiments, large areas are 467 covered by sea ice, even those that feature excessive mixed-layer depths (Fig.2). This is due to 468 the fact that in these simulations the water column does not contain water that is noticeably 469 warmer than the freezing point, i.e. excessive convection will not affect sea ice. After 40 years of 470 coupled integration with the high-resolution AGCM (TP6M-T255.40), the sea-ice cover and 471 thickness show a substantial increase (Fig.5) even though the deep-ocean accumulated a 472 considerable heat reservoir (Fig.9 left column). The increase in sea-ice extent and thickness is 473 thus due to a substantial reduction in vertical mixing, as reflected by a reduced mixed-layer depth

474 (Fig.11) and by the stronger stratification (Fig.9 right column). While the coupled simulation 475 with the coarser AGCM (TP6M-T63) led to a similar build-up of a heat reservoir in the deep 476 ocean, the mixed-layer thickness (not shown), even though somewhat reduced, remains 477 excessive and is still affecting wide regions 40 years into the coupled integration. Accordingly, 478 the stratification is weaker such that the sea ice can more readily be influenced by oceanic heat 479 flux. Fig.5 demonstrates that this goes to the point where open-ocean polynyas start emerging. 480 Thus, the higher resolution AGCM produces surface conditions that trigger a stronger 481 stratification.

482 In the following we shed more light on these relationships by examining in more detail why 483 the SO performance of TP6M improved upon coupling with ECHAM6, and that of GR15-OMIP 484 upon increasing the external freshwater flux. Both simulations end up with a substantially more 485 realistic strength of the ACC due to more realistic density south of the ACC. The improved ACC 486 in the forced GR15 experiments is exclusively caused by the enhanced external freshwater flux. 487 The ice cover and the ACC improving in concert with the HLSO water masses strongly suggests 488 that the additional freshwater flux is crucial for improving the model representation of the HLSO 489 in MPIOM. Considering further that the observed temperature and salinity profiles reflect a 490 rather complex water-mass distribution, maintaining it in a 50-year integration is remarkable. 491 GR15-OMIP-FW.50 also yields a substantial increase in deep-ocean salinity, and a substantial 492 salinity decrease in the upper 200 m, both changes leading to a rather realistic salinity profile. 493 This indicates a switch in the mode of operation from open-ocean deep convection to no 494 convection (e.g., Gordon et al., 2007; Martin et al., 2013), i.e., as long as there is convection, any 495 freshening of the surface leads to a freshening of the deep-ocean. Beyond a certain threshold of 496 surface freshening, stratification becomes strong enough to inhibit deep convection, which then 497 results in an immediate build-up of a heat and salt reservoir at depth, i.e. CDW, which is 498 ultimately maintained by the outflow of NADW (e.g., Pierce et al., 1995; Stössel and Kim, 2001). 499 The physical mechanism behind the direct relationship between freshwater supply in the 500 HLSO and HLSO density (below 200 m) seems to occur in two steps. First, surface freshening

increases the stratification, thus reducing the amount of dense-water formation through openocean convection. This allows for the build up of warm and saline CDW at intermediate depths.
At the same time, sea ice grows thicker and more compact (Fig.5), which reduces surface heat
loss in winter. This reduces ice growth and thus the amount of brine release, which in turn

reduces the amount of oceanic heat flux affecting sea ice, until a subtle balance establishes between a considerable heat reservoir at depth and a reasonably thick and compact ice cover at the surface that inhibits the production of excessively dense deep water by open-ocean convection. There is an obvious increase of ice thickness and extent with an increase in external freshwater supply, and correspondingly a considerable decrease in surface heat loss.

510 A similar change is noticed with the TP6M experiments when switching from uncoupled to 511 coupled mode, and when continuing the model integration in coupled mode. During the 40 years 512 of both coupled TP6M integrations, the strength of the ACC experiences a considerable decrease 513 toward observed estimates (Table 1; Fig.3). Besides changes in the external freshwater flux, this 514 change can in the coupled case also be due to a decrease in the strength of the southern 515 hemisphere westerlies from NCEP values to the ones provided by ECHAM6. Fig.12 shows the 516 meridional distribution of 6-year-mean zonal-mean zonal wind stress over the ocean south of 517 50°S for selected experiments (color) and observed climatologies/reanalyses (black). Switching 518 from NCEP.2010 to TP6M-T255.40 is associated with a substantial reduction of westerly wind 519 stress and with a northward shift of its core, both of which may readily explain the reduction of 520 the ACC (e.g., Fyfe et al., 2007). Another aspect related to convection in the HLSO is the 521 strength of the WS and RS gyre circulation, which is largely determined by the differential 522 Ekman transport and thus the wind-stress curl. A steeper meridional gradient in zonal wind stress, 523 such as NCEP relative to TP6M-T255 (Fig.12), will lead to a stronger WS gyre, which in turn 524 will lead to doming of isopycnals in its center and thus a more favorable environment for open-525 ocean convection. This mechanism has also been proposed in Cheon et al. (2014) who relate this 526 to the occurrence of the WS polynya in the 1970s. A more negative wind-stress curl, as e.g. 527 associated with a steepening of the zonal wind-stress curves in Fig.12 would therefor promote 528 open-ocean convection, a depletion of the heat reservoir at depth, an increase of HLSO 529 subsurface density, and thus eventually lead to an increase in ACC strength.

Accordingly, Table 1 also shows the 6-year mean wind-stress curl over the ocean region south of 50°S. For the experiments discussed so far there is a clear relationship between the slopes in Fig.12 and the magnitude of negative wind-stress curl. TP6M-NCEP.2010 clearly shows the largest negative wind-stress curl ( $-1.68 \cdot 10^{-7}$  Pa/m) and also the largest ACC strength. Immediately upon coupling, the curl diminishes to  $-1.27 \cdot 10^{-7}$  Pa/m and seemingly stays at that value over the next 40 years of coupled integration. The decrease of the ACC from 230 Sv to 141 536 Sy could thus be entirely due to the combined effect of reduced zonal wind stress and reduced 537 wind-stress curl. This is particularly supported by the fact that the total external freshwater flux 538 south of 60°S decreases upon coupling, thus promoting the preconditioning for convection. On 539 the other hand, if the freshwater flux south of  $70^{\circ}$ S is decisive (see section 3), its near doubling 540 and presumed northward advection will contribute to reduced convection. In any case, the 541 coupled simulations experience a considerable decrease of water density below 200 m south of 542 the ACC (Fig.4 and Table 1), which leads to a major reduction of the meridional density gradient 543 across the ACC, and thus a weakening of the ACC itself. It is important to note that this dramatic 544 reduction is seemingly just due to changes in the surface boundary conditions of the HLSO, i.e. 545 reduction in zonal wind stress and magnitude of negative wind-stress curl, surface freshening, 546 and warming upon coupling. All factors enhance the stratification of the ocean, reducing upward 547 oceanic heat flux (e.g., Bitz et al., 2005; Zhang, 2007; Goosse et al., 2009; Kirkman and Bitz, 548 2011; Martin et al., 2013), allowing the buildup of thicker ice (Fig.4) and a heat reservoir at depth (Fig.9 left column). Fig.10 shows the meridional winter sections along 30W of GR15-549 550 OMIP-FW.50 corresponding to Figs.1 and 2. The simulated vertical-meridional water-mass 551 distribution of this forced experiment is in close agreement with the corresponding WOCE 552 section (Fig.1), even in absolute values. TP6M-T255.40 shows a similar distribution (Fig.11), but 553 with much colder water below 2000 m depth. This is likely due to the deep part of the HLSO not 554 having adjusted after 40 years of coupled integration, as demonstrated for TP6M-T63.90 (Fig.9 555 left column, solid light blue line).

556 This study is only to examine the short-term ( $\sim$ 50 years), transient response of the HLSO to 557 varying surface boundary conditions. As can be seen when going from TP6M-T63.40 to TP6M-558 T63.90, 50 years is not sufficient time for the regional baroclinic part of the circulation of the SO 559 to adjust. Long-term impacts in association with the global overturning circulation were not 560 considered. On a time scale of several hundred years, the HLSO will be affected by the global 561 circulation, in particular by the properties and the rate of NADW outflow (e.g., Pierce et al., 562 1995; Drijfhout et al., 1996; Stössel et al., 2002; Martin et al., 2013). The long-term impact 563 emerges in the HLSO properties of the preindustrial control simulation of CMIP5, e.g. MPI-564 ESM-LR. This coupled simulation (in the current setting named GR15-T63-piCon) shows some 565 improvement over TP6M-NCEP.2010 (Fig.2), e.g. the existence of a substantial heat reservoir in 566 the central WS gyre (Fig.9 left column, marked purple line). In combination with excessive

567 convection, however, this yields much thinner ice than in the forced case to the point that open-568 ocean polynyas emerge (Fig.5). Furthermore, the deep ocean salinity of the central WS gyre is 569 substantially underestimated, with a concomitant underestimation of density (Fig.9 middle and 570 right column, marked purple lines). This misrepresentation is also evident from Table 1, GR15-571 T63-piCon showing the lowest HLSO density of all listed experiments, while the external 572 freshwater flux into the HLSO is lower than that of the TP6M coupled experiments. On the other 573 hand, the strength of the ACC turns out to be on the high side. From Table 1 and Fig.12 it is also 574 evident that the stronger ACC cannot be explained by differences in zonal wind stress and wind-575 stress curl. In this context it is interesting to note that the HLSO of the coupled simulation of the 576 corresponding NCAR model (CCSM4) is too salty by about 0.2 g/kg (Weijer et al., 2012), with a 577 concomitant excessive ACC strength of 173 Sv which seems to be mainly due to an excessive westerly wind stress of up 0.21  $N/m^2$  (Meijers et al., 2012). 578

579 A problem with the improved open-ocean water-mass characteristics in the HLSO of the 580 forced experiments is that the rate of AABW formation slows down to unacceptable values (e.g., 581 decrease of AABW intrusion into Atlantic from about 4 Sv to 1 Sv in GR15-OMIP-FW.50). A 582 similar reduction of AABW intrusion occurs in the coupled TP6M simulations, though from 9 Sv 583 to 5 Sv, which is acceptable (e.g., Ganachaud and Wunsch, 2000; Sloyan and Rintoul, 2001). In 584 these simulations, sea-ice thickness is overestimated along most of the Antarctic coastline, in 585 particular along east Antarctica, and coincides with large local freshwater input through runoff. 586 Excessive runoff may stabilize the water column to the point that not enough AABW is being 587 produced through the process of near-boundary convection. Considering that glacial melt water 588 spreads to a large extent through icebergs (e.g., Gladstone et al., 2001), it seems sensible and 589 physically justified to distribute part of the runoff from Antarctica over the HLSO, instead of 590 dumping it all into coastal grid cells. Such measure should suppress open-ocean convection and 591 simultaneously enhance near-boundary convection (e.g., Martin and Adcroft, 2010). To mimic 592 this effect in a most simplistic way, we conducted a sensitivity experiment with the coupled 593 GR15-T63 (MPI-ESM-LR) model starting from the preindustrial control year 1850, and 594 assuming that all Antarctic runoff (glacial meltwater) normally entering the coastal ocean grid 595 cells be distributed evenly over the ocean region south of  $60^{\circ}$ S, thereby conserving freshwater. 596 This provides an upper bound estimate of the effect of iceberg drift, as in reality about half of the

597 glacial melt water still enters the ocean right at the coast in the form of basal melt water598 (Depoorter et al., 2013).

599 The results are shown under GR15-T63-RO.50 in Table 1 and in most of the figures. As 600 expected, the total external freshwater flux south of 70°S (Table 1) and generally along the coast 601 (Fig.6) diminishes relative to GR15-T63-piCon (which would be the reference case for this set of 602 experiments), while the freshwater flux south of  $60^{\circ}$ S remains approximately the same. 603 Associated with the reduced freshwater flux south of  $70^{\circ}$ S is a higher sea-surface salinity (Fig.8). 604 The difference in coastal salinity is about 0.3 g/kg and extends down to about 300 m depth (not 605 shown). Accordingly, the density along the coast increases to much more realistic values 606 (Table1, Fig.4), suggesting that sea-ice formation and AABW formation along the coast have 607 improved. The larger high-latitude density may partly explain the stronger ACC (from 160 Sv to 608 178 Sv) (Table1; Fig.3) following our earlier explanation. Furthermore, the sea level in the 609 central WS declined by 3 cm, and the horizontal volume-transport streamfunction of the WS 610 increased by 2%, in line with the enhanced negative wind-stress curl over the HLSO (Table 1, 611 Fig.12). While the stronger ACC strength is less realistic compared to GR15-T63-piCon, there is 612 a slight improvement of the HLSO water-mass properties between 60°S and 70°S and in the 613 central WS (Fig.9 from marked purple profiles to plain solid purple profiles), indicating that a 614 longer integration may eventually lead to more realistic HLSO properties. The major 615 improvement occurring along the coast suggests that it is rather the reduction of freshwater flux 616 along the coast that leads to improvements than its increase further offshore, quite in line with 617 the findings of Martin and Adcroft (2010).

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## 620 5. Summary and Conclusions

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It can generally be concluded that MPIOM is capable of simulating the presently observed (e.g., Gordon et al., 2007) complex water-mass structure of the central WS provided the surface fluxes are such that they prevent open-ocean deep convection. A prerequisite for a realistic simulation is sufficient external freshwater input through precipitation or glacial melt water (e.g., Marsland and Wolff, 2001), a realistic westerly wind stress and wind-stress curl over the HLSO, and/or a more realistic redistribution of freshwater by sea ice (e.g., Haumann et al., 2014a). 628 Observed estimates on the amount and distribution of glacial melt water bear large uncertainties 629 (e.g., Gladstone et al., 2001; Beckmann and Goosse, 2003; da Silva et al., 2006; Jongma et al., 630 2009; Tournadre et al., 2012; Depoorter et al., 2013). Distributed over the ocean area south of 631  $60^{\circ}$ S, these estimates convert to about 10 cm/year (Table 1). Over the same area, the net annual 632 freshwater flux due to precipitation from reanalyses ranges from 60 to 70 cm/year (e.g., 633 Bromwich et al., 2011). Evaporation rates from reanalyses in this region seem more inconsistent, 634 probably due to differences in the way sea-ice concentration and -thickness are being prescribed. 635 We estimate the mean evaporation rate to be around 10 cm/year (e.g., Kållberg et al., 2005), thus 636 roughly compensating the estimated glacial melt water input to this region. Accordingly, the total 637 external freshwater flux of GR15-OMIP-FW.50 (Table 1) is overestimated by about 22 cm/year. 638 A problem with this simulation is that it is uncoupled so that buoyancy loss due to negative 639 sensible heat flux in wintertime is exaggerated. In order to arrive at reasonable HLSO water-640 mass properties, this can apparently be compensated by buoyancy gain from extra freshwater 641 flux.

642 Forced model experiments with the Large-Scale Geostrophic model (LSG) (Pierce et al., 643 1995) suggest that a freshwater input south of 60°S of about 82 cm/year is needed to support 644 regime shifts from a HLSO convective to non-convective mode of operation, and vice versa. 645 Martin et al. (2013), employing the coupled ECHAM5-NEMO model, achieve this with a 646 freshwater input of only about 47 cm/year (apparently, this value of precipitation minus 647 evaporation includes glacial melt/continental runoff). As can be seen from Table 1, the HLSO 648 freshwater flux value of Pierce et al. (1995) corresponds to our uncoupled experiment GR15-649 OMIP-FW.50, whereas the lower value of Martin et al. (2013) is closer to those of our coupled 650 experiments TP6M-T255 and TP6M-T63. These experiments showing substantial improvements 651 of HLSO water-mass properties in comparison with TP6M-NCEP.2010, an addition of 10 652 cm/year south of 60°S from about 55 cm/year to observed estimates of about 65 cm/year in the 653 coupled TP6M experiments will likely be sufficient for shutting down open-ocean convection. 654 This is somewhat counterintuitive as TP6M-NCEP.2010 has apparently more realistic external 655 freshwater flux values than any of the coupled TP6M experiments, but then it features also the 656 strongest westerly wind stress and strongest negative wind-stress curl.

657 An issue not analyzed in this study is the internal freshwater flux, i.e. that associated with 658 the annual rate and pattern of sea-water freezing and sea-ice melting (e.g., Bitz et al., 2005; 659 Kirkman and Bitz, 2012; Haumann et al., 2014a). Wind field errors will have a direct impact on 660 sea-ice drift and the pattern and rate of annual net freezing (e.g., Holland and Kwok, 2012; Close 661 and Goosse, 2013; Turner et al., 2013; Holland et al., 2014). Another uncertainty arises from the 662 way sea-ice salinity is being treated. In MPIOM, it is fixed at a value of 5. This is at the low end 663 of possible values for first-year ice (e.g., Vancoppenolle et al., 2009; Hunke et al., 2011), which 664 is the dominant ice type in the SO. A higher sea-ice salinity of about 8 would diminish brine 665 release during sea formation, and thus deliver a stabilizing buoyancy flux in fall and winter in the 666 open ocean, but also along the coast. In any case, insufficient sea-ice export from the coastal 667 regions will contribute to underestimating near-boundary convection and to overestimating open-668 ocean convection. This role of the sea-ice induced freshwater flux has been investigated by 669 Haumann et al. (2014a) based on satellite data.

670 With respect to the question of whether using a higher-resolution (and thus substantially 671 more expensive) atmosphere model in coupled simulations delivers more realistic results in the 672 HLSO the answer based on this study is yes. TP6M-T255 delivers more realistic winter sea ice 673 and water-mass properties than TP6M-T63, while the ACC strength is equally well represented. 674 Considering that the coupling frequency is identical, differences are solely due to horizontal 675 resolution. Pronounced differences between TP6M-T255 and TP6M-T63 occur in wind stress 676 right along Antarctica, in particular along Victoria Land (165<sup>0</sup>E), which is orographically the 677 extension of the Transantarctic Mountain range. These differences resemble those found by 678 Stössel et al. (2011) in this region. They compared T799 (i.e. about 20 km) resolution wind fields 679 from the ECMWF operational analysis with those of the NCEP reanalyses, which have a 680 resolution of about 200 km (about T63) in this region. A more detailed resolution of the coastal 681 winds is crucial for the model representation of coastal polynyas and thus the rate of AABW 682 formation through near-boundary convection. The reasons for discrepancies related to the 683 resolution of atmospheric fields will be the subject of further investigations.

As shown in this study, and consistent to earlier findings, the main players controlling the HLSO properties are the amount and distribution of the freshwater flux, and the strength and meridional gradient of the southern hemisphere westerlies. Continuing a forced eddy-resolving ocean simulation in coupled mode improves the HLSO properties considerably, at least over a time span of 40 years. Increased resolution of the atmosphere model leads to further improvements. The Antarctic runoff of the coupled simulations appears too concentrated along 690 coastal grid points. Relaxing this situation by distributing this runoff over a wider region 691 improves the HLSO water-mass properties, in particular along the coastline. This contributes to a 692 more realistic representation of AABW formation through near-boundary convection, while the 693 effect on open-ocean convection seems minor.

Recent literature on the performance of the HLSO focused on intercomparisons among CMIP5 models (e.g., Russell et al., 2006; Meijers et al., 2012; Zunz et al., 2013; Close and Goosse, 2013; Heuzé et al., 2013; Sallée et al., 2013; Turner et al., 2013). This study suggests that it may be worth scrutinizing the HLSO performance of individual ESMs under various conditions (model resolution, parameterizations, forced vs. coupled, initial conditions, glacial meltwater, transient vs. equilibrium conditions) to eventually arrive at more realistic simulations and a better understanding of this highly dynamic and complex region of the world's ocean.

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Experiments	Den- sity <70S	<i>FWF</i> <70S	Den- sity <60S	FWF <60S	P – E <60S	Run- off <60S	Resto -ring <60S	ACC	Wind stress curl
GR15-OMIP-REF.50	27.85	34	27.84	47	26	9	12	195	- 1.30
GR15-OMIP-FW.50	27.69	95	27.78	87	131+	9	- 53	142	- 1.30
TP6M-NCEP.2010	27.88	57	27.87	73	44	9	20	230	- 1.68
TP6M-T255.0	27.86	87	27.87	62	46	16	n/a	219	- 1.27
TP6M-T255.40	27.70	79	27.81	59	47	12	n/a	141	- 1.28
TP6M-T63.40	27.70	80	27.81	54	41	13	n/a	148	- 1.52
TP6M-T63.90	27.70	80	27.81	56	43	13	n/a	135	- 1.59
GR15-T63-piCon	27.64	61	27.71	47	37	11	n/a	160	- 1.43
GR15-T63-RO.50	27.71	38	27.73	47	38	9	n/a	178	- 1.49
Observations*	27.81	55	27.81	65	55	10	n/a	135	

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714 Table 1: Annual-mean as well as south-of-70°S and south-of-60°S spatial-mean

715 density below 200 m depth (referenced to surface pressure)  $[kg/m^3-1000]$  and

716 external (excluding contributions from sea-ice growth or melt) freshwater (FW)

717 flux [cm/year], consisting of precipitation (P), evaporation (E), runoff, and FW flux

from restoring when uncoupled, as well as ACC strength [Sv] and wind-stress curl [Pa/m \* 10-7] over the ocean south of 50°S from all simulations considered

719 curl [Pa/m \* 10<sup>-7</sup>] over the ocean south of 50°S from all simulations
720 and as derived from observations.

721 \*) Some of the observed estimates bear large uncertainties (explanations and 722 references see text).

<sup>+)</sup> This figure includes specified additional freshwater flux (FW experiment).





Fig.1: Meridional sections along 30°W of potential temperature (left), salinity (right), and potential density (contours) from the WOCE climatology.









732 Fig.3: Time series of annual-mean Drake Passage throughflow from uncoupled 733 experiments GR15-OMIP experiments (colored dashed lines) REF (dark green) 734 and FW (light green), and from coupled experiments (colored solid lines) TP6M-735 T255 (orange), TP6M-T63 (dark blue), and GR15-T63-RO (purple).



738 Fig.4: August-mean, vertical-mean density below 200 m for indicated

739 experiments and WOCE.





742 Fig.5: 5-year August-mean sea-ice thickness for indicated experiments.





745 Fig.6: Annual-mean external freshwater flux for indicated experiments.



Fig. 7: August-mean 2-m air temperature (color) and sea-ice thickness (white
lines, increment: 0.3 m, range: 0 to 0.6 m) for indicated experiments.





52 Fig.8: August-mean uppermost model layer salinity for indicated experiments.



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Fig.9: Potential-temperature (left), salinity (middle), and potential-density (right) 756 757 profiles at 30°W, 65°S (top), in region 20°-40°W, 60°-70°S (middle), in region 0°-360°E, 60°-70°S (bottom) from WOCE climatology (solid black), from uncoupled 758 759 GR15-OMIP experiments (colored dashed lines) REF.50 (dark green), FW.50 (light green), and from coupled experiments (colored solid lines) TP6M-T255.0 (red), 760 TP6M-T255.40 (orange), TP6M-T63.40 (dark blue), TP6M-T63.90 (light blue), 761 762 GR15-T63-RO.50 (purple), and GR15-T63-piCon (purple with symbols). All 763 simulation results reflect August-mean conditions.













770 Fig.11: Results from TP6M-T255.40; otherwise as Fig.2.



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- 773

774 Fig. 12: 6-year-mean, zonal-mean wind stress from coupled experiments (colored

775 solid lines) TP6M-T255.40 (orange), TP6M-T63.40 (dark blue), GR15-T63-RO.50

776 (purple), and GR15-T63-piCon.50 (purple with symbols), as well as year 2010

777 NCEP (solid black) and OMIP (dashed black).

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