

# Parameterized internal wave mixing in three ocean general circulation models

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## Key Points:

- the IDEMIX parameterization for internal wave mixing is evaluated in three global ocean models using three different tidal forcing
- in agreement with observations, simulations using IDEMIX show stronger and spatially more variable mixing patterns than those without IDEMIX
- most circulation and water mass changes can be attributed to stronger mixing but some effects are likely blurred by numerical mixing

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## 17 Abstract

18 The parameterization IDEMIX for vertical mixing by breaking internal gravity waves  
 19 is evaluated in three different non-eddy resolving ocean models. Three different prod-  
 20 ucts of wave forcing by tidal flow over topography, representing the current uncertainty,  
 21 are applied and compared to reference simulations without IDEMIX, allowing the model-  
 22 independent effects of the new closure to be assessed. Common to all models is larger  
 23 interior mixing work with stronger horizontal structure due to the inhomogeneous forc-  
 24 ing functions in all simulations using IDEMIX, in better agreement to observations. Co-  
 25 herent model responses to the stronger mixing work are changes in the thermocline depth  
 26 including IDEMIX related to stronger shallow overturning cells in the Indo-Pacific Ocean.  
 27 Furthermore, deeper mixed layer depths in the subpolar North Atlantic are related to  
 28 an increase of the Atlantic overturning circulation which brings the model closer to ob-  
 29 servations, coming along with an increase in northward heat transport. In the South-  
 30 ern Ocean, excessive energy input by one of the forcing products leads to unrealistic deep  
 31 convection in the Weddell Sea in one of the models. The deep Indo-Pacific overturning  
 32 circulation and the bottom cell of the Atlantic feature an incoherent model response, which  
 33 may point towards the importance of excessive numerical mixing in the models.

## 34 Plain Language Summary

35 Gravity waves propagate not only at the ocean’s surface but also in the ocean in-  
 36 terior. These internal gravity waves are generated mostly at the bottom by oscillating  
 37 tidal flow across topographic obstacles. Similar to the surface waves, the internal waves  
 38 can break and mix the surrounding water when the waves get shorter but their ampli-  
 39 tude remains the same. Such interior wave breaking mixes dense water upwards which  
 40 is important to drive large-scale flows such as the global overturning circulation of the  
 41 world’s ocean.

42 Current ocean models cannot resolve the small-scale wave breaking. Therefore, this  
 43 important effect needs to be implemented by a parameterization. In this study, we use  
 44 the new parameterization IDEMIX which is based on internal wave dynamics and en-  
 45 ergetics. This parameterizations is implemented in three different ocean models. A nec-  
 46 essary input for the IDEMIX parameterization is the energy input into the wave field  
 47 at the bottom of the ocean. In this study, we use three different products for this energy  
 48 input to study the sensitivity of the simulated circulation with respect to this forcing dif-  
 49 ferences. Common model responses are diagnosed and discussed, while some incoherent  
 50 model responses regarding the overturning circulation pose new questions.

## 51 1 Introduction

52 In the ocean’s meridional overturning circulation, turbulent mixing is responsible  
 53 for the return of abyssal dense water to the surface through the isopycnals, even though  
 54 there is little turbulent energy available to drive substantial diapycnal flows (Munk &  
 55 Wunsch, 1998). A lot of this mixing is thought to be associated with breaking internal  
 56 waves. Internal waves are generated by, for instance, tidal or geostrophic flows over to-  
 57 pography or by fluctuating winds (Olbers, 1983; Polzin & Lvov, 2011). While they prop-  
 58 agate through the ocean they can become subject to non-linear wave-wave interactions  
 59 through which energy is transferred to waves with smaller and smaller wave lengths un-  
 60 til the waves become unstable, break and produce small-scale turbulence (see e.g. Müller  
 61 et al., 1986). Unfortunately, the life cycle of an internal wave involves a large range of  
 62 space and time scales, so that its observation or simulation poses many challenges (Müller  
 63 et al., 1986). As a result, large uncertainties remain regarding the understanding and  
 64 quantification of single aspects of this life cycle. Furthermore, many aspects of internal  
 65 wave dynamics are not resolved in global ocean models used in state-of-the-art climate

66 models and will most likely not be resolved in the foreseeable future. Instead, the effect  
 67 of the internal wave breaking and the associated diapycnal mixing needs to be param-  
 68 eterized in these ocean models to account for the important driving mechanism of di-  
 69 apycnal flows.

70 Owing to the complexity of the problem and the sparsity of adequate observations,  
 71 interior internal wave driven mixing is often treated rather simply in many ocean mod-  
 72 elling endeavours. In particular, vertical mixing parameters are often chosen without tak-  
 73 ing into account physical or energetic constraints with respect to the sources of mixing.  
 74 They are rather treated as tuning coefficients to optimize certain aspects of the respec-  
 75 tive model simulations. For some vertical mixing schemes like the PP (Pacanowski & Phi-  
 76 lander, 1981) or KPP (Large et al., 1994) scheme it is common practise to let the ver-  
 77 tical diffusivity fall back in the interior to often spatially constant background values of  
 78  $O(10^{-5} \text{ m}^2/\text{s})$ . An analogous approach in higher order mixing closures like the TKE scheme  
 79 of Gaspar et al. (1990) is to impose an unphysically motivated minimal (constant) tur-  
 80 bulent energy. The basic assumption behind both choices is that the internal wave field  
 81 supplies a certain but unknown amount of energy to turbulent mixing, generating either  
 82 the turbulent energy or the mixing rate itself. However, it is obvious that both approaches  
 83 are not physically consistent with the dynamics of internal waves, and will not consis-  
 84 tently represent the observed spatio-temporal variability of wave-induced turbulent mix-  
 85 ing. Motivated by observations of enhanced mixing rates close to rough topography, Simmons  
 86 et al. (2004) attempt to include this variability by using an ad-hoc length scale for the  
 87 vertical shape and a map of tidal energy forcing for the horizontal distribution, to con-  
 88 struct a three-dimensional field of mixing rates. However, this closure remains heuristic  
 89 without considering wave physics, and the wave energy cycle is not treated consis-  
 90 tently.

91 The Internal wave Dissipation, Energy and Mixing (IDEMIX) model (Olbers & Eden,  
 92 2013) is a parameterization framework built to consistently account for the internal wave  
 93 physics. IDEMIX includes processes such as an energy flux into the internal wave field  
 94 by tides, surface winds, and other forcing functions, horizontal and vertical propagation  
 95 and refraction of internal waves, wave-wave interaction, wave-mean flow interaction, and  
 96 the conversion of internal wave energy to small-scale turbulence associated with wave-  
 97 breaking. The IDEMIX model, however, depends crucially on the forcing of tides and  
 98 waves, but the magnitude and spatial pattern of these forcing functions bear large un-  
 99 certainties. One aim of this study is thus to estimate the uncertainty of key aspects of  
 100 the ocean circulation rooted in the uncertainty of the tidal forcing. To this end, we com-  
 101 pare three simulations with different tidal forcing products applied to IDEMIX with ref-  
 102 erence simulations without IDEMIX, in which small-scale turbulence is parameterized  
 103 by a constant minimum background value. The different forcing products are derived  
 104 from (1) a scaling law for internal tide generation applied in barotropic ocean models  
 105 using a bulk wave number for topography (Jayne & St. Laurent, 2001), (2) a direct cal-  
 106 culation from linear theory applying a more realistic bottom topography and using eight  
 107 tidal constituents from a tidal model (Nycander, 2005; Falahat et al., 2014) and (3) es-  
 108 timates of internal tide generation from a high-resolution ocean model (Li & von Storch,  
 109 2020) complemented with higher constituents from linear theory. The results are eval-  
 110 uated with respect to water mass biases, circulation changes, and mixing rates obtained  
 111 from observations.

112 The effect of different parameterizations is often model dependent. To assess this  
 113 effect, we apply three different representative state-of-the-art ocean models: ICON-O (Korn  
 114 et al., 2022), FESOM (Danilov et al., 2017), and MITgcm (Marshall et al., 1997). The  
 115 models are very similar in their implementation of IDEMIX, share the same surface forc-  
 116 ing, and are similar albeit not equal in their vertical and horizontal resolution. The mod-  
 117 els also have substantial differences: most importantly, ICON and FESOM use (differ-  
 118 ent) triangular grids in the horizontal, while the MITgcm uses a classical rectangular grid.

**Table 1.** Most important features of the numerical models used in this study. Note that the effective horizontal resolution is difficult to compare on the different grids. Here we simply give the nominal grid spacing.

	ICON-O	FESOM	MITgcm
horizontal resolution	ca. 40 km	ca. 20–100 km	ca. 20–111 km
vertical levels	64	48	50
grid type	triangular	triangular	rectangular
grid staggering	C-grid	B-grid	C-grid

119 ICON and the MITgcm use a C-grid discretization, while FESOM uses a B-grid. A complete  
 120 description of similarities and differences of the three models is beyond the scope  
 121 here, the reader is referred to the key references of the models given here and below. In  
 122 any case, we aim to differentiate between model-independent effects of the IDEMIX closure  
 123 with different forcings functions and the model-dependent effects by including three  
 124 different models.

125 In the following section 2, we detail the model setups and parameter choices. In  
 126 Section 3, we discuss the effect of the different internal wave forcing products in the different  
 127 models on the mixing work and compare to available observations of mixing in Section 4. In  
 128 Section 5 the simulated water masses and in Section 6 the impact on the circulation are discussed.  
 129 Section 7 provides discussion and conclusion.

## 130 2 Numerical model configurations and experiments

131 We use in this study three different numerical models with similar configurations.  
 132 These models are the MITgcm, FESOM and ICON-O. All configurations have been developed  
 133 for other studies which do not include IDEMIX, and all model parameters are chosen according  
 134 to their respective default values obtained from the previous general model performance tuning.  
 135 Here we only unify the vertical mixing parameterizations in all models without retuning the  
 136 models. Some important model features are listed in Tab. 1. In all models, meso-scale eddies  
 137 are not resolved but parameterized by a bolus velocity (Gent et al., 1995) and isopycnal  
 138 diffusion (Redi, 1982). ICON uses a constant thickness mixing coefficient, FESOM a  
 139 vertically varying coefficient following Ferreira et al. (2005), and the MITgcm simulation  
 140 uses a horizontally varying coefficient based on horizontal and vertical buoyancy gradients  
 141 (Visbeck et al., 1997). Furthermore, all three models differ in the numerical implementation  
 142 of the parameterization (Korn, 2018). The MITgcm and FESOM simulations use a vertical  
 143  $z^*$ -coordinate (Adcroft & Campin, 2004) while ICON uses  $z$ -levels. All models use a  
 144 non-linear free surface. More details about the specific model configurations can be found  
 145 for MITgcm, FESOM and ICON-O in Forget et al. (2015), Scholz et al. (2022), Korn et al.  
 146 (2022), respectively.

147 All simulations are forced by the same wind stress, and surface heat and freshwater  
 148 fluxes are computed with the same bulk formulae (Large & Yeager, 2009) from atmospheric  
 149 fields of the 1958–2019 Japanese Re-Analysis dataset JRA55-do-v1.4.0 (Tsuji  
 150 no et al., 2018). For all simulations, the models are integrated for five consecutive forcing  
 151 cycles of 62 years. In addition to applying freshwater fluxes, surface salinity is relaxed  
 152 towards its initial condition with a piston velocity of  $10 \text{ m}/60 \text{ days} = 0.1666 \text{ m d}^{-1} =$   
 153  $1.929 \times 10^{-6} \text{ m s}^{-1}$ . Initial conditions for temperature and salinity are also identical  
 154 and derived from January values of the PHC-3.0 climatology (Steele et al., 2001). If not  
 155 stated otherwise, we diagnose time averages over the last 40 years (1979–2019) of the last  
 156 forcing cycle. While the total integration time of 310 years might be too short for the sim-

157 uations to reach an equilibrium, it is still sufficiently long to study the major implica-  
 158 tions of vertical mixing on water masses and circulation.

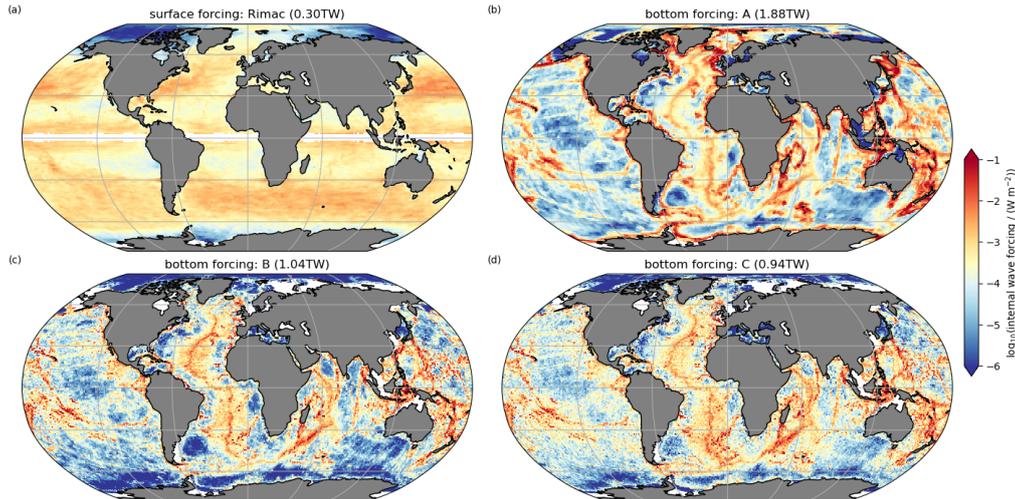
159 For each model, we discuss four different experiments with different sources of internal  
 160 wave energy available for mixing. In our reference configuration vertical mixing  
 161 is parameterized by the higher order turbulent kinetic energy (TKE) closure by Gaspar  
 162 et al. (1990) and Blanke and Delecluse (1993), without making use of IDEMIX (exper-  
 163 iments FESOM-REF, ICON-REF, and MITgcm-REF, respectively). Here, the diapyc-  
 164 nal interior mixing is determined by resetting the turbulent kinetic energy to a mini-  
 165 mum background level. This approach implicitly assumes that an unknown internal wave  
 166 field always provides this level of energy for mixing in the ocean interior. A detailed de-  
 167 scription of the closure by Gaspar et al. (1990) is given in Appendix A. In the other three  
 168 experiments carried out with each model, we include the IDEMIX closure (Olbers & Eden,  
 169 2013), which predicts the propagation and dissipation of the wave energy, and the mini-  
 170 mum background level of turbulent kinetic energy of the original TKE-scheme is accord-  
 171 ingly set to zero. A detailed description of the IDEMIX closure is given in Appendix B.

172 Our version of IDEMIX requires surface and bottom wave forcing, given by the near-  
 173 inertial wind-driven surface pumping and by interaction of barotropic tides with topog-  
 174 raphy, respectively. In the sensitivity experiments with IDEMIX, the surface forcing re-  
 175 mains the same, but we apply three different forcing datasets for the bottom forcing to  
 176 reflect the current uncertainty of internal wave generation by the tides: the forcing as  
 177 described in Jayne and St. Laurent (2001) (FESOM-A, ICON-A, MITgcm-A), the forc-  
 178 ing derived from linear theory after Nycander (2005) (FESOM-B, ICON-B, and MITgcm-  
 179 B), and the forcing derived to large parts from the STORMTIDE2 simulation (Li & von  
 180 Storch, 2020) (FESOM-C, ICON-C, and MITgcm-C)

181 In the following, we only list the most important features of these three bottom forc-  
 182 ing datasets and refer to a more detailed description in Appendix C. Forcing A (Fig. 1b)  
 183 is based on a scaling law for internal tide generation (and barotropic tide dissipation)  
 184 suggested by Jayne and St. Laurent (2001) that is motivated by linear theory (Bell, 1975b)  
 185 and used in barotropic tidal models to represent the drag exerted by the baroclinic on  
 186 the barotropic tide. The energy flux by this drag diagnosed from barotropic tidal mod-  
 187 els is often used in the heuristic parameterization of Simmons et al. (2004), for exam-  
 188 ple, in the CESM model (Hurrell et al., 2013). One parameter in the scaling law is the  
 189 bulk wavenumber of the bottom topography, which is treated as a tuning parameter in  
 190 the barotropic tidal models. This means that the effect of bottom topography on the gen-  
 191 eration of internal tides may not be represented very accurately in forcing A, that is, forc-  
 192 ing A is subject to (unknown) errors of the barotropic model.

193 To avoid this source of errors, one can alternatively derive the bottom forcing di-  
 194 rectly from linear theory and realistic bottom topography at high resolution. Here we  
 195 use the estimates of Nycander (2005) as calculated by Falahat et al. (2014) (forcing B  
 196 hereafter, Fig. 1c), which takes as input the barotropic velocities from a tidal model us-  
 197 ing the eight major tidal constituents and the observed topographic spectrum. Both forc-  
 198 ings A and B are, however, subject to the limitations of linear theory. For example, lin-  
 199 ear theory breaks down for topographic slopes steeper than the internal tide beam.

200 Estimating the tidal bottom forcing from internal tide generation in ocean general  
 201 circulation models that are forced by the full tidal potential avoids issues of linear the-  
 202 ory entirely. As an example for this method, we derive forcing dataset C (Fig. 1d) from  
 203 a concurrent simulation of circulation and tides by the Max Planck Institute Ocean Model  
 204 referred to as STORMTIDE2 (see Li & von Storch, 2020, for details of the model setup  
 205 and the computation of the internal tide generation). Restrictions of linear theory do  
 206 not apply in such simulations, but the finite horizontal resolution (about  $0.1^\circ$ ) allows only  
 207 the first few vertical internal wave modes to be excited, and the parameterization of dis-  
 208 sipation may introduce additional unknown model errors. In addition, the conversion of-

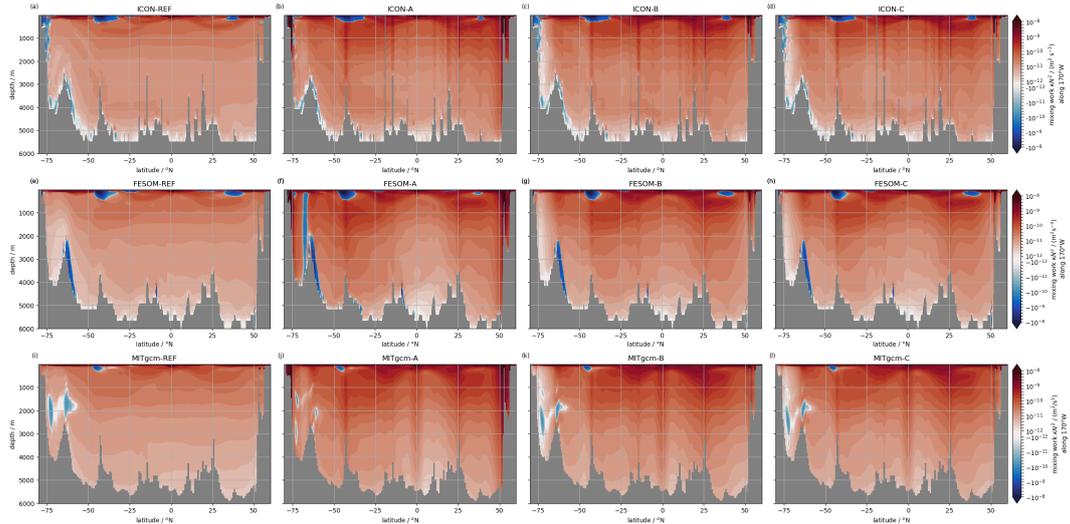


**Figure 1.** Energy flux into the internal wave field mapped to the ICON grid from (a) wind-driven near-inertial surface pumping (Rimac et al., 2013) and the bottom (tidal) forcings A-C (b-d). See text for more details.

209 ten becomes negative, which is not necessarily unphysical (Kelly & Nash, 2010) but means  
 210 it cannot be used directly as a (per definition positive) forcing term in IDEMIX. We here  
 211 follow the procedure of de Lavergne et al. (2019) to remove negative values while pre-  
 212 serving the original depth-dependent conversion rate. The model simulation by Li and  
 213 von Storch (2020) includes the full luni-solar tidal forcing. However, only the internal  
 214 tide generation by the  $M_2$  tide was calculated, so that the other seven constituents of  
 215 the computation by Nycander (2005) are added to complete the forcing C. The model  
 216 simulation includes full luni-solar tidal forcing. However only the internal tide genera-  
 217 tion by  $M_2$  tide was calculated. These seven constituents account for roughly a third of  
 218 of the globally integrated conversion in forcing C. In summary, all three bottom forcing  
 219 datasets have their own limitations and it is unclear a-priori which has the lowest biases  
 220 and what the implications are on simulated circulation and watermass structure if the  
 221 forcing is applied in a numerical model.

222 All tidal forcing datasets have in common that the energy flux is enhanced over ma-  
 223 jor topographic obstacles like sea mounts and ridges, for example, along the Mid-Atlantic  
 224 Ridge. The forcings B and C are in general smaller in magnitude than forcing A, espe-  
 225 cially in the Southern Ocean, and the global integral of forcing A (1.9 TW) is about two  
 226 times larger than for B (1.0 TW) and C (0.9 TW). Note that the estimate of forcing B  
 227 excludes supercritical slopes where linear theory breaks down and thus excludes depths  
 228 above 400 m. In contrast, the forcing A includes also the continental shelves (Fig. 1b).  
 229 In the model-based forcing C, there are no large conversion rates at the shelf break. This  
 230 may point towards a bias of the barotropic models used to generate forcing A. A discus-  
 231 sion of the reasons for the differences of the forcing datasets is beyond the scope of the  
 232 current study; here we consider the differences as plausible error bounds for the bottom  
 233 forcing.

234 Oscillations in the horizontal divergence of wind driven currents in the surface mixed  
 235 layer with frequencies at or larger than the local Coriolis frequency generate downward  
 236 propagating internal waves at the base of the mixed layer. This process is called (near-  
 237 inertial) wind-driven surface (Ekman) pumping (Olbers et al., 2020; von Storch & Lüs-  
 238 chow, 2023). For all simulations, we use the same associated energy flux into the internal wave



**Figure 2.** Mixing work  $\kappa N^2$  along a meridional section at  $170^\circ\text{W}$ . First row (a-d) shows results from ICON, second row (e-h) those from FESOM and third row (i-l) those from the MITgcm.

239 field derived from the global estimate of Rimac et al. (2013) (Fig. 1a) since the global  
 240 integral (0.3 TW) is much smaller than the tidal forcing (Fig. 1b-d).

### 241 3 Simulated mixing work

242 The main source of small-scale turbulence in the interior ocean is due internal wave  
 243 breaking. Thus, the internal wave forcing controls the interior turbulent kinetic energy  
 244 that is available for mixing of water mass properties. Note that in our reference exper-  
 245 iments the internal wave breaking is parameterized by resetting small turbulent kinetic  
 246 energy values to an arbitrary constant minimum, but in the simulations with IDEMIX,  
 247 this source of energy is parameterized based on physical principles. There are two ma-  
 248 jor sinks of turbulent energy in the interior ocean: molecular dissipation with subsequent  
 249 transformation into heat, and upward mixing of dense water masses, which represents  
 250 the transformation of turbulent kinetic to mean potential energy. The upward energy  
 251 flux is given by  $\kappa N^2$ , with the diapycnal diffusivity  $\kappa$  and the square of the buoyancy  
 252 frequency  $N$ . In the following, we will refer to  $\kappa N^2$  as mixing work.

253 Fig. 2 shows the mixing work  $\kappa N^2$  along  $170^\circ\text{W}$  in the different experiments. In  
 254 the surface mixed layer,  $\kappa N^2$  is low or even negative, but in the interior  $\kappa N^2$  is mostly  
 255 positive. In the Southern Ocean close to the bottom,  $\kappa N^2$  is negative in all ICON and  
 256 FESOM simulations ( $N^2$  gets small or negative there, not shown). In FESOM-A, a large  
 257 region with  $\kappa N^2 < 0$  in the Southern Ocean extends even towards the surface. This  
 258 is related to a potentially too strong wave forcing which effects also the mixed layer depth  
 259 and circulation, and which will be further discussed below. In the MITgcm simulations,  
 260 there are small and negative values of  $\kappa N^2$  at mid depth in the Southern Ocean (Fig. 2i-  
 261 l) associated with the circulation in the Ross Sea. These features can be traced back to  
 262 a feedback between vertical mixing and stratification in regions of strong horizontal flow:  
 263 generally salinity stabilizes the water column and temperature destabilises it in this re-  
 264 gions. When horizontal advection and vertical mixing reduces the small vertical salin-  
 265 ity gradients they can no longer stabilize the water column leading to further mixing.

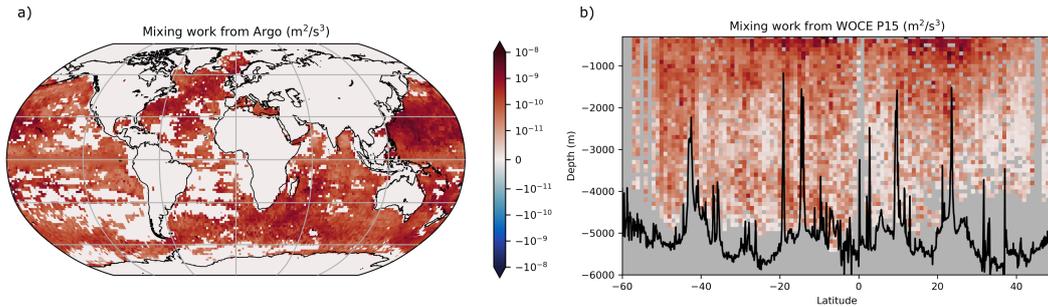
266 In general, the magnitude of the mixing work in the ocean interior is similar in all  
 267 experiments:  $\kappa N^2$  decreases from  $10^{-8} \text{ m}^2/\text{s}^3$  close to the surface to  $10^{-11} \text{ m}^2/\text{s}^3$  at depth.  
 268 In all experiments with IDEMIX, however,  $\kappa N^2$  tends to be larger compared to the re-  
 269 spective reference experiment, in particular towards the surface. In all reference exper-  
 270 iments, the parameterized homogeneous source of turbulence in the experiments ICON-  
 271 REF, FESOM-REF, and MITgcm-REF yields a relatively homogeneous mixing, whose  
 272 structure is mainly shaped by  $N^2$ . In contrast, the simulations including IDEMIX with  
 273 different tidal forcings show a much richer horizontal structure with larger mixing work  
 274 due to enhanced wave forcing over rough bathymetry. Differences for the individual mod-  
 275 els between the experiments with different tidal forcings are smaller than the difference  
 276 to the respective reference experiment. The experiments with forcing A tend to have slightly  
 277 larger  $\kappa N^2$  than those using the other forcings in accordance with the larger energy in-  
 278 puts, but all agree roughly in the location and magnitude of the mixing hot spots.

279 For the simulations including IDEMIX, the total energy available for mixing is the  
 280 globally integrated internal wave forcing from Fig. 1, which amounts to 2.13, 1.27, and  
 281 1.18 TW for forcing A, B, and C, respectively (taking bottom and surface forcing together).  
 282 For the reference simulations without IDEMIX, this available mixing energy is param-  
 283 eterized as the change of energy (per time) that is necessary to keep the turbulent ki-  
 284 netic energy at its depicted background value which in our case is  $1 \times 10^{-6} \text{ m}^2 \text{ s}^{-2}$ . In-  
 285 tegrating this rate of change leads to a mixing work of 0.37, 0.28, and 0.25 TW for ICON-  
 286 REF, FESOM-REF, and MITgcm-REF, respectively. We interpret these values as an  
 287 energy supply to the interior turbulent kinetic energy, which can be compared with the  
 288 total wave forcing in the simulations including IDEMIX. The mixing work in the refer-  
 289 ence simulations is thus by far lower than in the simulations including IDEMIX and of  
 290 a similar magnitude as the wind induced forcing. In principle, we could increase the mix-  
 291 ing work in the reference simulations by choosing a different background value for tur-  
 292 bulent kinetic energy, but by doing so the observed horizontal structure with its mix-  
 293 ing hot spots will not be reproduced in the reference simulation. Therefore, we keep the  
 294 commonly used background parameter and make no attempt to change it.

295 The diapycnal diffusivities  $\kappa$  are shown in Fig. C1. Horizontal variations of  $\kappa$  are  
 296 stronger when IDEMIX is applied comparable to the mixing work (Fig. 2). One excep-  
 297 tion is ICON-REF, where also an enhanced horizontal structure can be found, this struc-  
 298 ture is accompanied by a similar structure in  $N^2$  (not shown) such that the product  $\kappa N^2$   
 299 is smooth (Fig. 2a). In all simulations,  $\kappa$  is enhanced in the upper ocean mixed layer.  
 300 In ICON and FESOM, enhanced diffusivities can also be found at the bottom and in the  
 301 Southern Ocean (in particular in FESOM-A, where the  $170^\circ\text{W}$  section cuts through the  
 302 deep convection area in the Weddell Sea of this simulation). In the MITgcm simulations,  
 303 the diffusivities are also enhanced in the Southern Ocean between 1000 m and 3000 m  
 304 in accordance to the unstable conditions which occur in the simulations of this model  
 305 (as discussed above). In general, all simulations with IDEMIX have higher diffusivities  
 306 in accordance with the higher amount of energy available for mixing.

## 307 4 Comparing mixing work with observation

308 Direct observations of small-scale turbulent mixing are sparse. Therefore, we com-  
 309 pare our model simulations to indirect estimates that were obtained from hydrographic  
 310 profiles by applying the finestructure method (e.g. Gregg, 1989; Kunze et al., 2006; Polzin  
 311 et al., 2014). The finestructure method links small-scale turbulence to finescale internal  
 312 gravity wave variability based on a parameterization of wave energy dissipation by wave-  
 313 wave interactions. Note that this parameterization is also used in IDEMIX and given  
 314 by  $\epsilon_{iw}$  in Eq. B1. This form for  $\epsilon_{iw}$  is confirmed by numerical evaluation of the scatter-  
 315 ing integral for wave-wave interactions (Eden et al., 2019). The finestructure estimates  
 316 have a substantially larger uncertainty (a factor of three and more) than microstructure  
 317 or direct turbulence observations (Pollmann et al., 2017). We can therefore reliably com-

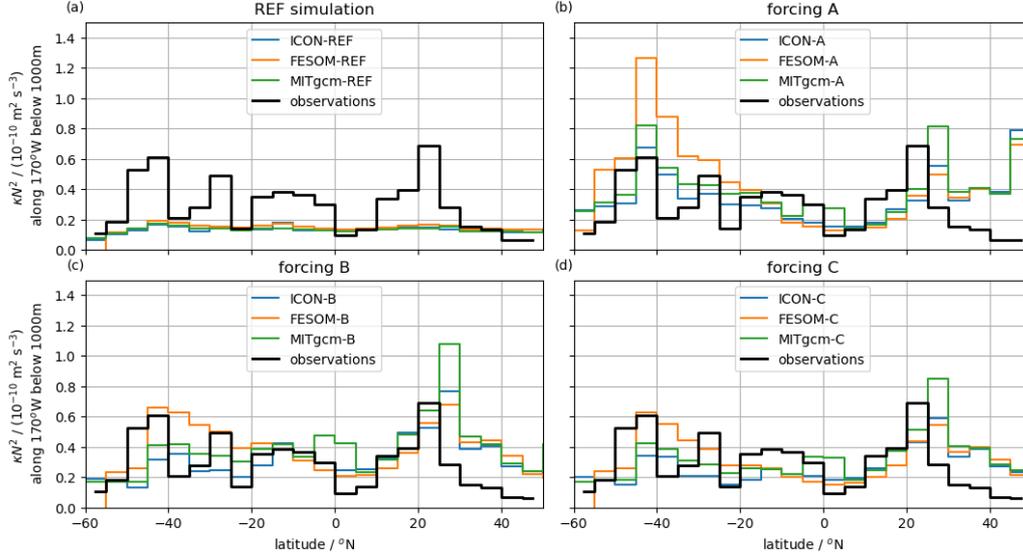


**Figure 3.** Mixing work  $\kappa N^2$  compiled from the finestructure estimates of (a) Pollmann et al. (2017) and (b) Kunze (2017). The data in (a) represent an average between 300 m and 2000 m depth. The data in (b) are averaged between 164.9°W and 165.1°W north of the equator and between 169.9°W and 170.1°W at and south of it (WOCE section P15). The black lines represent the topography from Becker et al. (2009) (SRTM30+), showing the 0 m-isobath in (a) and the bottom topography in (b).

318 pare the observed quantities to our model simulations only where variations in magni-  
 319 tudes are sufficiently large, that is, larger than the error bound of a factor of three. We  
 320 here use (a) an estimate of the vertical diffusivity and TKE dissipation rates from Argo  
 321 float profiles (Pollmann et al., 2017) and (b) a data base derived from the finestructure  
 322 method applied to WOCE/CLIVAR hydrographic sections (Kunze, 2017) (Fig. 3).

323 The mixing work  $\kappa N^2$  varies geographically in the global ocean in the depth range  
 324 accessible by Argo float observations (above 2000 m, Fig. 3a). The variations span sev-  
 325 eral orders of magnitude, with relatively low values along the equator and over the abyssal  
 326 plains and maxima near mixing hot spots associated with rough bottom topography (e.g.  
 327 the Hawaiian and Emperor Seamount Chains and the Izu-Bonin-Mariana arc system)  
 328 and eddy activity (e.g. the Gulf Stream and Kuroshio regions). The WOCE section P15  
 329 runs roughly along  $170 \pm 5^\circ$ W, so that we can compare the P15  $\kappa N^2$  estimate (Kunze,  
 330 2017) (Fig. 3b), with Fig. 2 (note the different  $x$ -axis limits owing to limited data avail-  
 331 ability). By definition, the finestructure method is only applied where  $N^2 > 0$ , so in  
 332 contrast to the model simulations, the mixing work contrary is always positive. The mix-  
 333 ing work decreases with depth from maximum values of  $10^{-8} \text{ m}^2 \text{ s}^{-2}$  near the surface to  
 334 minimum values of  $10^{-11} \text{ m}^2 \text{ s}^{-2}$  and less at intermediate depth and, in some locations,  
 335 near the sea floor. This vertical structure and the overall magnitude is similar in all model  
 336 simulations using IDEMIX.

337 In the horizontal, the observed mixing work shows four bands of elevated mixing  
 338 work ( $\kappa N^2$ ) almost for the entire water column. These are located south of 40°, around  
 339 30°S, around 20°N, and roughly between 20°S and the equator. These variations are sig-  
 340 nificant within the uncertainty of the finestructure method (Pollmann et al., 2017): When  
 341 averaging the mixing work in the water column below 1000 m (that is, below the observed  
 342 surface maxima and, in case of the model simulations, also minima) and in 5° latitude  
 343 bands, it is elevated by a factor of 6.5 (40°-45°S), 5.2 (25°-30°S) and 7.3 (20°-25°N) rel-  
 344 ative to the equatorial band at 0°-5°N. For any of the 5° latitude bands between the equ-  
 345 ator and 20°S, the corresponding factor is between 3 and 4. At these locations, topographic  
 346 features stand out from the otherwise small-scale roughness or almost plain surface of  
 347 the sea floor: Near 40°S, the WOCE P15-section crosses Chatham Rise of Zealandia, near  
 348 25°S Louisville Ridge, near 20°N the Hawaiian Island Chain and the surrounding seamounts,  
 349 and between the equator and roughly 20°S, there are the seamounts and islands of the  
 350 Samoan Basin (e.g. Harris et al., 2014; Mortimer et al., 2017). These topographic fea-



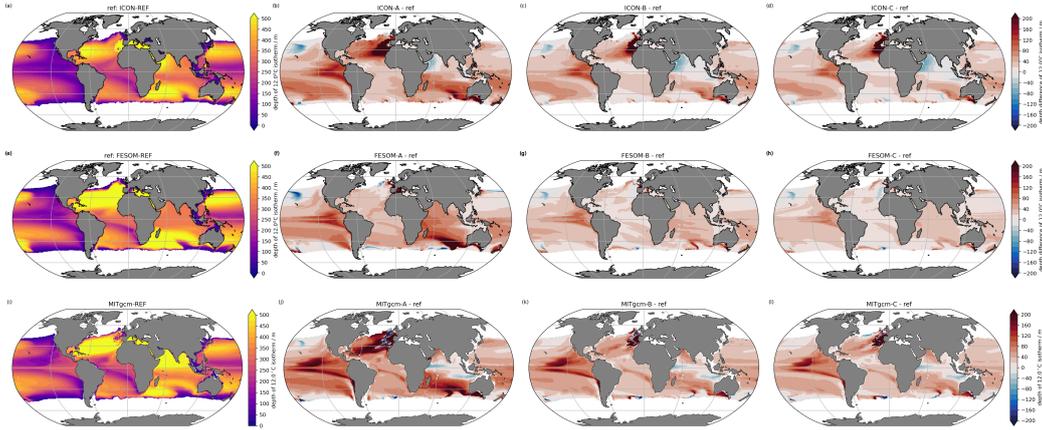
**Figure 4.** Mixing work along 170°W averaged below 1000 m for the reference simulation (a) and the IDEMIX simulations with forcing data A, B, and C in (b), (c), (d), respectively. The black line represents the corresponding results for the observed mixing work shown in Fig. 3b. All results are binned in 5° latitude intervals.

351 tures generate internal tides (Falahat et al., 2014, their Fig. 6), and, as some fraction of  
 352 the generated baroclinic tidal energy is dissipated locally (Vic et al., 2019, their Fig. 5),  
 353 hence the observed increase of mixing work  $\kappa N^2$ .

354 These characteristics are reproduced in ICON, FESOM, and MITgcm to a differ-  
 355 ent degree depending on the model and the forcing, but they are only reproduced when  
 356 IDEMIX is active (Fig. 4): Without IDEMIX, the variability of the modeled  $\kappa N^2$  along  
 357 170°W is almost negligible compared to the observed one (Fig. Fig. 4a) and the mod-  
 358 eled values are also notably lower. With IDEMIX, however, all models show a higher mix-  
 359 ing work and reproduce the minima at higher latitudes and near the equator as well as  
 360 the maxima at around 45°S and 20°N (Fig. 4b-d).

361 The increase in mixing work in high-mixing bands compared to the 0-5°N band is  
 362 up to a factor of 2 (MITgcm), a factor of 2-3 (ICON) and a factor of 2-3 (FESOM, reach-  
 363 ing a factor of 7 for forcing A) with IDEMIX, and below 30% in the reference case with-  
 364 out IDEMIX. The other two maxima of  $\kappa N^2$  seen in the observations are not reproduced  
 365 by the models, with exception of the ICON simulation, which features a third peak just  
 366 south of the equator for forcing B and forcing C. There might be several explanations  
 367 for this, one being that the forcing itself is not as strong as at the other two locations  
 368 where elevated mixing work is observed along 170°W (see Fig. 1).

369 Although the agreement between modeled and observed variability differs between  
 370 the different forcing setups, a robust conclusion as to which forcing data leads to the best  
 371 agreement with observations cannot be drawn owing to the relatively large uncertainty  
 372 of the finestructure method. Moreover, there is also some spread among the models, and  
 373 elevated mixing work south of the equator in FESOM for forcing A.



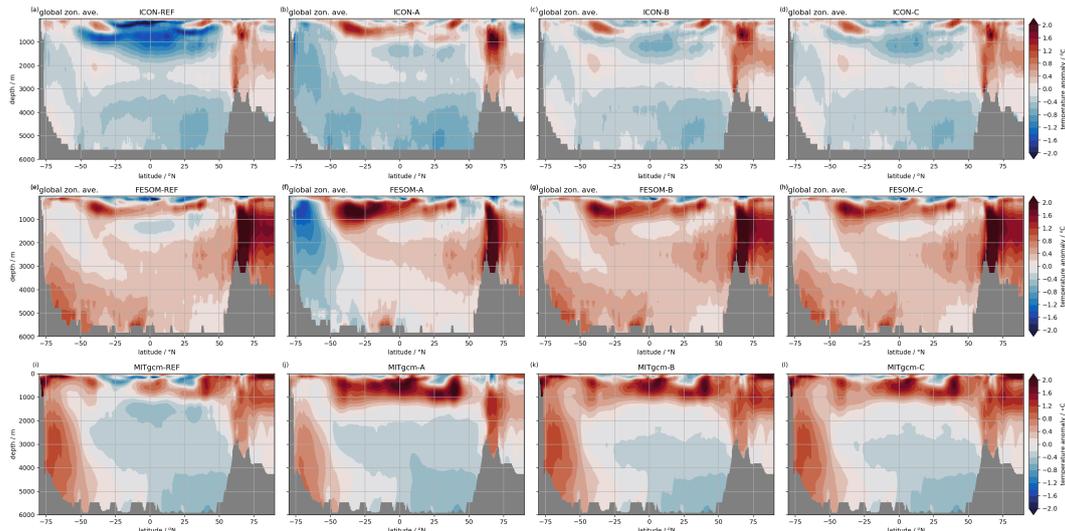
**Figure 5.** Depth of the  $12^{\circ}\text{C}$  isotherm as an indication of the thermocline depth. First row (a-d) shows results from ICON-O, second row (e-h) FESOM, and third row (i-l) MITgcm. The first column shows results for the respective reference simulation, the other three columns show differences to the respective reference simulation.

## 5 Effects on water masses

The different levels of energy available for mixing have implications for water mass transformations in the model simulations. We choose the depth of the  $12^{\circ}\text{C}$  isotherm as a proxy for the thermocline depth. More mixing work moves the thermocline downwards and less mixing lifts this isotherm upwards. With more mixing work available, all simulations that include IDEMIX have generally deeper thermoclines compared to the reference simulations (Fig. 5). The differences are not uniform and there are even locally shallower thermoclines with IDEMIX. For all three models, the deepest thermoclines can be found in the simulations with the strongest forcing A. Simulations with forcing B or C have comparable thermocline depths, in between those of the reference simulations and the simulations with forcing A. All models and all tidal forcings produce comparable differences of the thermocline depth when compared with the reference simulations. The strongest increase of the thermocline depth can be found in the eastern tropical Pacific, the eastern sub-tropical Atlantic and the southern Indian Ocean, but those areas are not necessarily related to enhanced tidal forcing and wave dissipation. The small regions of shallower thermocline depths also coincide between the different model simulations, showing a rather coherent model response in the thermocline to the change in vertical mixing.

Fig. 6 shows zonally averaged temperature biases compared to the initial conditions in the different simulations. While ICON-REF is too cold within the thermocline and too warm close to the surface within  $50^{\circ}\text{S}$  and  $50^{\circ}\text{N}$ , the other models are too warm within the thermocline and too cold at the surface in the reference simulations. The stronger mixing in the IDEMIX simulations changes these biases, since stronger upper-ocean mixing decreases surface temperatures and increase temperatures within the thermocline. For the ICON model, where thermocline waters are too cold in the reference simulation, the biases are reduced when IDEMIX is included (improvement is largest for ICON-A and less for ICON-B and ICON-C). For the other two models (FESOM and MITgcm), for which the thermocline waters are already too warm in the reference simulations, including IDEMIX increases these biases even further.

A prominent temperature bias which does not change with IDEMIX is located in the North Atlantic at  $50^{\circ}\text{N}$  (Fig. C2). All models with all forcing products are too warm

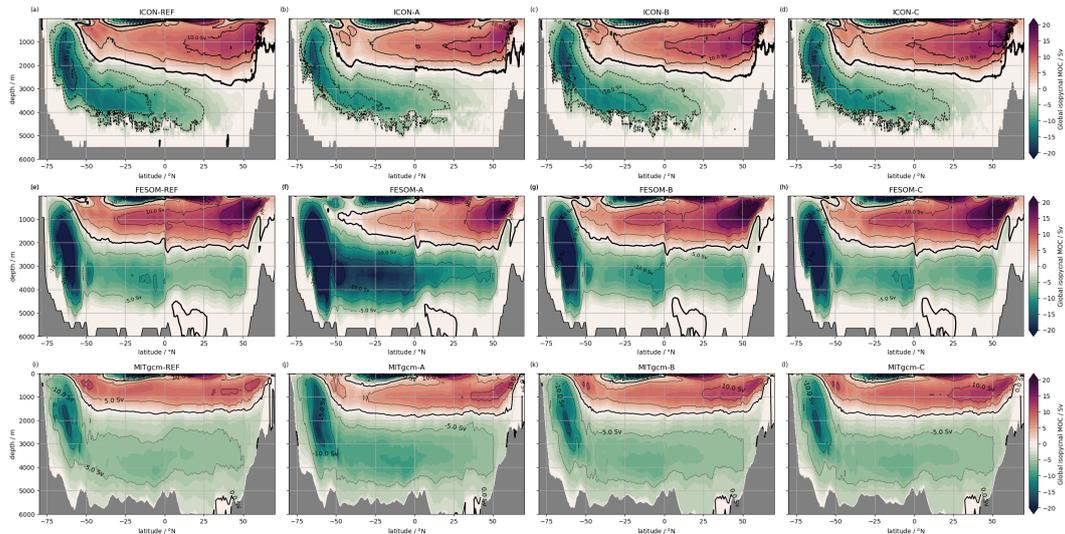


**Figure 6.** Zonal average of the temperature bias with respect to the observed initial conditions for ICON (a-d), FESOM (e-h) and MITgcm (i-l).

405 in this area and it appears that there are no fundamental changes when IDEMIX is applied.  
 406 This bias is related to the (missing) north-west corner of the North Atlantic Current,  
 407 whose dynamics is most likely related to (the interaction of) meso-scale eddies, topography,  
 408 and dense bottom flow. The horizontal model resolutions do not allow to resolve mesoscale eddies,  
 409 and the eddy closure used in the models is not designed to account for the effect of eddy-topography  
 410 interaction. Since the overflow areas further north are neither horizontally nor vertically well resolved,  
 411 a low bias in the dense bottom water flow might also contribute to the missing north-west corner in the models.  
 412 The warm bias at 50°N in the North Atlantic is most likely unrelated to biases in vertical mixing,  
 413 and consequently using IDEMIX does not change much the model results here.  
 414

415 The mixed layer depths in the subpolar North Atlantic (supplementary Fig. C4),  
 416 are increased in all models and in all experiments with IDEMIX compared to the respective  
 417 reference simulations, in particular in the convectively active regions such as the Nordic  
 418 Seas, Irminger, and Labrador Sea. This increase could be caused by a more efficient pre-  
 419 conditioning for convection in case of stronger vertical mixing in the convective regions.  
 420 It is largest with forcing A and smaller but similar for the other two forcing datasets.  
 421 The deeper mixed layer depths are associated with an increase of the meridional over-  
 422 turning circulation in the Atlantic Ocean as discussed in the next section. In MITgcm-  
 423 REF and ICON-REF, the mixed layer depths in the subpolar North Atlantic roughly  
 424 agree with observations (Fig. C6, Locarnini et al., 2018; Zweng et al., 2019), while FESOM-  
 425 REF tend to feature too deep convection depths here. In MITgcm-A und ICON-A, mixed  
 426 layer depths are getting too deep in comparison with the observations, but using forcing  
 427 B and C the region of deep mixing is increasing and tend to be in better agreement  
 428 with the observations. In FESOM, the experiments with IDEMIX are increasing the already  
 429 too deep and too large mixing region, thus increasing the model bias further.

430 In FESOM, mixed layer depths in the Southern Ocean also tend to be too deep compared  
 431 to observations, and also increase in the experiments with IDEMIX compared to  
 432 FESOM-REF, in particular in FESOM-A (supplementary Fig. C5). Related to the un-  
 433 realistically deep mixed layer in FESOM-A, the warm bias in the Southern Ocean in FESOM-  
 434 REF changes into a cold bias in FESOM-A (Fig. 6). Since forcing A has much larger am-  
 435 plitudes in the Southern Ocean and in particular in the Weddell Sea than the other two



**Figure 7.** Global meridional overturning stream function  $\psi$  in Sv.  $\psi$  was calculated in density space and remapped to depth levels for ICON (a-d), FESOM (e-h) and MITgcm (i-l).

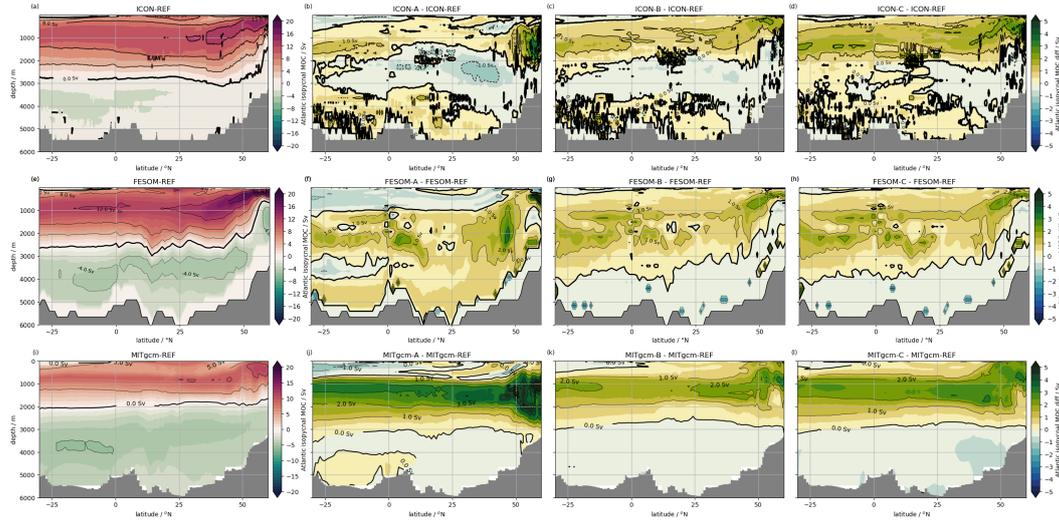
436 forcing datasets, it seems that the preconditioning by stronger mixing leads to the ex-  
 437 aggerated deep convection in FESOM-A. This is also related to a substantially enhanced  
 438 formation of bottom water in FESOM-A discussed in the next section. In ICON-A, we  
 439 also see slightly increased mixed layer depth in the Weddell Sea, which may point to-  
 440 wards too large forcing by dataset A in this region, but the other experiments with ICON  
 441 and also the MITgcm show hardly any changes.

## 442 6 Effects on the circulation

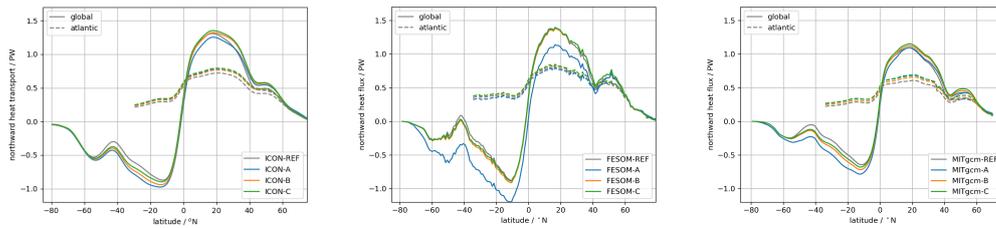
443 The global meridional overturning stream function  $\psi$  was calculated by zonally and  
 444 vertically integrating the northward transports below more than 80 isolines of the local  
 445 potential density  $\sigma_2(z)$  with reference pressure of 2000 dbar, and remapping to depth us-  
 446 ing  $z(\bar{\sigma}_2)$ , where  $\bar{\sigma}_2$  denotes the zonal average of  $\sigma_2$ . Note that this procedure reflects  
 447 better the actual watermass transports than simply averaging the transports at constant  
 448  $z$ -levels (McDougall & McIntosh, 2001). All simulations show the familiar two cell struc-  
 449 ture of the global overturning (Fig. 7). The differences in the strength and shape of the  
 450 overturning cells are larger between different models than between the different exper-  
 451 iments with each model (with the exception of FESOM-A).

452 While ICON and FESOM show in general stronger and deeper reaching upper cells  
 453 in the northern hemisphere, MITgcm has weaker overturning there. The lower cell in the  
 454 southern hemisphere is also weaker in MITgcm and stronger in ICON and FESOM. The  
 455 largest difference within the experiments with each model is given by the large increase  
 456 of the bottom cell in FESOM-A. In this FESOM simulation, the stronger mixing work  
 457 of forcing A seems to trigger exaggerated deep water formation in the Southern Ocean  
 458 that also leads to the mixed layer depth bias in Fig. C5. As a consequence, the over-  
 459 turning of the deep Pacific cell increases substantially. Except for the deep cell in the South-  
 460 ern Ocean, differences in  $\psi$  by different forcings remain relatively small, compared to the  
 461 differences between the models.

462 Also for the decomposition of the global stream function into Atlantic and Indo-  
 463 Pacific basins, the differences between the models are larger than differences between dif-  
 464 ferent experiments with the same model (Fig. C7 and C8 in the appendix). In all mod-



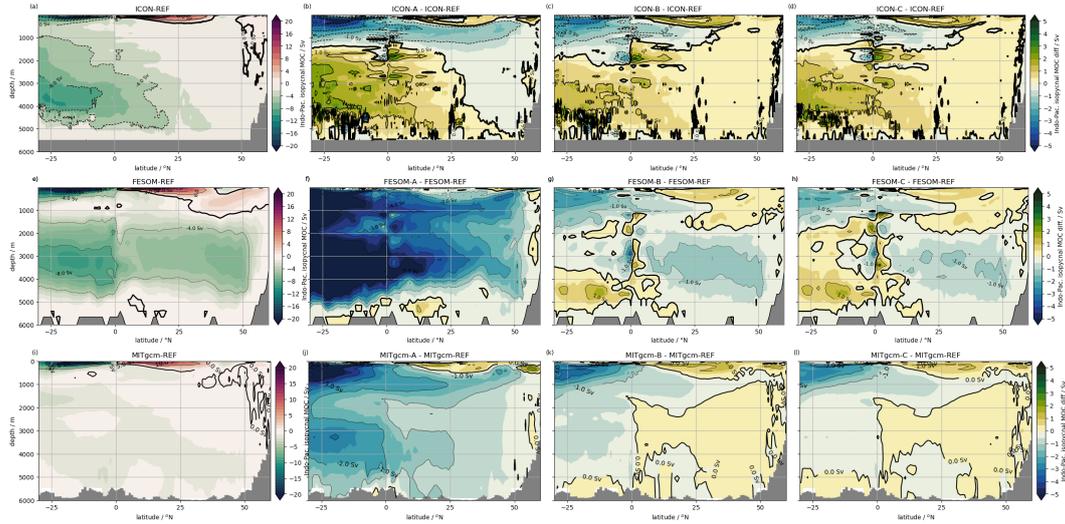
**Figure 8.** As Fig. 7, but for the Atlantic basin, and showing differences to the reference experiments to the sensitivity experiments with different forcings.



**Figure 9.** Northward heat transport in PW.

465 els, the Atlantic upper cell increases by up to 5 Sv, but the vertical shape of this increase  
 466 is different between models (Fig. 8). For ICON and FESOM, the increase of the upper  
 467 cell in the subtropics is largest for forcing C, but for the MITgcm is is for forcing A. The  
 468 increased overturning is related to deeper mixed layers in the subpolar North Atlantic  
 469 in each of the experiments with IDEMIX (compare supplementary Fig. C4), which points  
 470 towards increased deep water formation. This relation between deep convection and the  
 471 strength of the overturning is often seen in ocean models (e.g. Eden & Jung, 2001). How-  
 472 ever, the connection between deep water formation and overturning is still not fully under-  
 473 understood (e.g. Lozier et al., 2019).

474 Compared to observations (e.g., Lumpkin & Speer, 2007) the upper cell of the At-  
 475 lantic overturning is too weak in all reference simulations, in particular in the subtrop-  
 476 ics and in MITgcm-REF. Furthermore, the upper cell of the stream function is too shal-  
 477 low (see Korn et al., 2022; Jungclaus et al., 2022). The common model response to in-  
 478 cluding IDEMIX thus tends to drive all models closer to observations. Accordingly, north-  
 479 ward heat transports are also increased in the Atlantic Ocean (compare Fig. 9), but all  
 480 models still fall short in reproducing the observed (more than 1 PW) heat transports.  
 481 Changes in the lower cell in the Atlantic are weak and incoherent between the different  
 482 models (Fig. 8), a feature also seen in the bottom cell of the Indo-Pacific, discussed be-  
 483 low.



**Figure 10.** Same as Fig. 8, but for the Indo-Pacific basin.

484 In the Indo-Pacific basins, the strength of the southern upper shallow overturning  
 485 cell within the thermocline increases in all models with IDEMIX compared to the refer-  
 486 ence simulations (Fig. 10). North of the equator, a similar increase of the shallow cell  
 487 is also seen in ICON and MITgcm, but only to a weaker extent in FESOM. The shal-  
 488 low cells within the thermocline are thought to be wind-driven, therefore the common  
 489 model response of an increase of the cells due to larger mixing work is surprising. How-  
 490 ever, this increase is related to a deeper thermocline in these simulations (Fig. 5), which  
 491 might lead to larger areas of the subducting density layers exposed to the atmosphere  
 492 and thus stronger ventilation and stronger overturning.

493 Substantial changes can be found in the Indo-Pacific bottom cell (Fig. 10), but no  
 494 coherent response amongst the models, similar to the Atlantic bottom cell. The largest  
 495 change in the Indo-Pacific can be seen in FESOM-A, which we can relate to the exag-  
 496 gerated deep convection in the Weddell Sea in this simulation. MITgcm-A also shows  
 497 an increase in the bottom cell due to the stronger mixing in the Southern Ocean com-  
 498 pared to the very weak cell in MITgcm-REF, but in contrast to our expectation, the bot-  
 499 tom cell is decreasing in strength in all ICON simulations including IDEMIX and the  
 500 larger mixing work. A bottom water transport estimate of  $22.7 \pm 7.7$  Sv into the Pacific  
 501 at  $32^\circ$ S from observations (Lumpkin & Speer, 2007) is not reproduced by any model, even  
 502 for the unrealistic case FESOM-A (other observational estimates yield smaller transports  
 503 into the Pacific). The reason for this model bias, and in particular the reason for the in-  
 504 coherent model response in the bottom cell in the Indo-Pacific and Atlantic Ocean re-  
 505 mains unclear. We discuss this aspect in the next section.

## 506 7 Discussion and conclusions

507 A vertical mixing scheme based on internal wave physics (IDEMIX) is implemented  
 508 and evaluated in three different ocean model codes: ICON-O, FESOM, and MITgcm.  
 509 The implementation of IDEMIX is available at [https://github.com/nbruegge/CVMix-  
 510 src](https://github.com/nbruegge/CVMix-src). The implemented IDEMIX version (Olbers & Eden, 2013) predicts the bulk wave  
 511 energy propagation and dissipation, given the wave forcing functions at the top and the  
 512 bottom. The wave energy dissipation provides forcing to an energy-based mixing closure  
 513 for the effects of small-scale turbulence (Gaspar et al., 1990), and is available for mix-  
 514 ing. The surface energy forcing for IDEMIX is small and we keep it the same in all model

515 experiments, while we apply three different products for the bottom forcing by tidal flow  
 516 over topography: forcing A is based on the drag parameterization by internal tide gen-  
 517 eration of a barotropic tidal model (Jayne & St. Laurent, 2001), forcing B is calculated  
 518 from linear theory using only velocities from a barotropic tidal model and the observed  
 519 topography spectrum (Nycander, 2005; Falahat et al., 2014), and forcing C is taken from  
 520 a global high-resolution ocean model simulation including tidal forcing (Li & von Storch,  
 521 2020). While forcing A is subject to the biases of the barotropic tidal model for which  
 522 the drag parameterization accounts for, forcing B suffers from the limitation of linear  
 523 theory of shallow slopes and weak flow, and forcing C from limited horizontal resolution,  
 524 dissipation and other unknown biases of the high-resolution model. Accordingly, the forc-  
 525 ing functions differ by almost a factor of two in the global integrated flux into the wave  
 526 field, where forcing A is the strongest, and B and C are similar, and represent the cur-  
 527 rent uncertainty in the flux into the wave field.

528 The three ocean models applied here are taken as examples for typical state-of-the-  
 529 art non-eddy resolving ocean-only global configurations. The surface forcing of the mod-  
 530 els is identical, while many other aspects of the models differ; the reader is referred to  
 531 the references listed in the main text about details of the model configurations. It is im-  
 532 portant to note that no attempt has been made to tune the performance of the new ver-  
 533 tical mixing scheme. The effect of the three different bottom forcing functions in the three  
 534 ocean models is assessed by comparing to a reference simulation, in which the effect of  
 535 breaking internal gravity waves is implemented by a threshold for minimal turbulent en-  
 536 ergy. Since the effect of new parameterizations is often model dependent, the common  
 537 response in the three different models allows us to assess the model-independent effects  
 538 of the IDEMIX closure. We find the following common model response:

- 539 • Common to all models is larger interior mixing work  $\kappa N^2$  in the global integral  
 540 with larger vertical and in particular horizontal structure due to the inhomoge-  
 541 neous forcing functions in all simulations using IDEMIX, compared to the respec-  
 542 tive reference simulation (Fig. 2). The global underestimation in the reference sim-  
 543 ulations could be resolved by adjusting the threshold of minimal turbulent energy  
 544 in the scheme of small-scale turbulent mixing, but the spatial structure will not  
 545 be reproduced without IDEMIX. Note that our choice of threshold value is based  
 546 on common practice in ocean modelling.
- 547 • IDEMIX improves the horizontal variations of  $\kappa N^2$  along  $170^\circ\text{W}$  – within the large  
 548 error bounds – compared to to finestructure observations. Forcing A seems to over-  
 549 estimate  $\kappa N^2$  in the South Pacific and the Southern Ocean and to underestimate  
 550 it in the subtropical Pacific. The mixing work obtained with forcing B best matches  
 551 the observations. All models but MITgcm tend to overestimate  $\kappa N^2$  in the sub-  
 552 polar North Pacific. However, the differences of  $\kappa N^2$  for the different forcing func-  
 553 tions stay within the large error bounds of the observations.
- 554 • A rather coherent model response are the changes in the thermocline depth. In  
 555 all simulations with IDEMIX the thermocline tends to become deeper compared  
 556 to the respective reference simulations, although there are also regions with shal-  
 557 lower thermocline depths (Fig. 5). This is related to cooling of the upper thermo-  
 558 cline and warming of the lower thermocline, but local thermocline depth changes  
 559 are not necessarily related to locally enhanced mixing rates. Whether these changes  
 560 drive the models closer to observations (i.e., the initial conditions) turns out to  
 561 be model dependent, since it may or may not compensate other model biases.
- 562 • The wind-driven shallow overturning cell in the Indo-Pacific within the thermo-  
 563 cline increases in all models with IDEMIX (Fig. 10). Due to the deepening of the  
 564 thermocline, larger areas of the subducting density layers might be exposed to the  
 565 atmosphere which then leads to the stronger ventilation and stronger overturn-  
 566 ing.

- 567 • Another common model response are deeper mixed layer depths in the subpolar  
568 North Atlantic, which could be due to more efficient preconditioning of deep con-  
569 vection (Fig. C4). In the Southern Ocean, the energy input in forcing A leads to  
570 an unrealistic large region of deep convection in the Weddell Sea in one of the mod-  
571 els, which is not or to a lesser extent present using the other forcing functions in  
572 the same model (Fig. C5). This artifact points towards too large and unrealistic  
573 energy input by forcing A (see above and Fig. 4b)
- 574 • The increase in mixed layer depth in the subpolar North Atlantic is related to a  
575 common increase of the upper cell of the Atlantic overturning circulation in all mod-  
576 els, driving the models closer to observed transports in the subtropics (Figs. 8 and 10).
- 577 • The increase in the upper cell of the Atlantic overturning circulation is associated  
578 with an increase in northward heat transport in the Atlantic in all models (Fig.  
579 9), although all models still underestimate northward heat transports.
- 580 • The lower cell in the Atlantic and the Indo-Pacific do not show systematic changes  
581 common to all models although changes are up to a factor of two in some of the  
582 simulations (Figs. 8 and 10). The lower Indo-Pacific cell in ICON weakens when  
583 IDEMIX is included which is counter intuitive with the enhanced mixing work present  
584 in the simulation applying IDEMIX. FESOM and MITgcm show an increase of  
585 the lower Indo-Pacific cell for the stronger bottom forcing A but hardly any change  
586 for forcing B and C. Furthermore, all simulations show too low transports of the  
587 lower Indo-Pacific cell.

588 The reason for the circulation bias in the bottom cell of the Indo-Pacific Ocean,  
589 and the reason for the incoherent model response in the bottom cell in the Indo-Pacific  
590 and Atlantic Ocean remains unclear to us. This incoherent model response is surpris-  
591 ing since it is commonly assumed that the bottom cells in the major ocean basins are  
592 driven by vertical mixing. On the other hand, the upper cell in the North Atlantic shows  
593 a coherent model response of an increase with stronger mixing work, although it is com-  
594 monly assumed that it is driven by wind stress in the Southern Ocean and not by the  
595 vertical mixing. The increase in the upper cell in the North Atlantic is related to deeper  
596 convection in the subpolar North Atlantic, which we in turn explain by changes in pre-  
597 conditioning for convection by the change in vertical mixing. We cannot answer how changes  
598 in convection are related to changes in the strength of the upper cell in the North At-  
599 lantic, since there is currently no consistent dynamical framework of the dynamics of the  
600 ocean's overturning in closed basins (e.g. Straub, 1996; Greatbatch & Lu, 2003; Brüggemann  
601 et al., 2011).

602 Strong numerical mixing typical for coarse models may hide some of the effects the  
603 explicit vertical mixing by IDEMIX on the large-scale transports in the bottom overturn-  
604 ing cells in the major ocean basins. Other, non-local effects may be responsible for the  
605 model biases in the bottom cells, such as deep water formation biases around the Antarc-  
606 tic, errors in bottom topography, or errors in the isopycnal structure of the Antarctic Cir-  
607 cumpolar Current. Unfortunately, numerical mixing is difficult to assess. We are work-  
608 ing on implementing methods to diagnose numerical mixing following Klingbeil et al. (2014)  
609 into the models, but for now we have to postpone a further discussion.

610 In any case, we could show that applying a more realistic vertical mixing param-  
611 eterizations have a notable effect on the ocean circulation with partly improved model  
612 biases. The more realistic mixing parameterizations furthermore help to identify model  
613 biases since the energy available for vertical mixing is finally physically constrained. One  
614 suspicious candidate for such a model bias is numerical mixing and the results which we  
615 obtained here suggest to more carefully revisit water mass transformations and diapyc-  
616 nal velocities associated with this numerical mixing.

## Appendix A Vertical mixing closure

In this study, we use the closure by Gaspar et al. (1990) to parameterize the mixing in the surface mixed layer, but also the mixing in the interior of the ocean in the reference experiments. The closure is based on a parameterized budget for turbulent kinetic energy  $E_{tke}$ , assuming lateral homogeneous conditions, given by

$$\partial_t E_{tke} = \partial_z c_{tke} k_m \partial_z E_{tke} + k_m (\partial_z \mathbf{u})^2 + \epsilon_{iw} - k N^2 - c_\epsilon E_{tke}^{3/2} L^{-1} \quad (\text{A1})$$

with the parameter  $c_\epsilon = 0.7$  and  $c_{tke} = 30.0$ .  $\epsilon_{iw}$  denotes the dissipation of internal wave energy and is defined below in Eq.B1. Key for the closure by Gaspar et al. (1990) is the mixing length assumption for the vertical viscosity  $k_m = c_k E_{tke}^{1/2} L$  with  $c_k = 0.1$ , or vertical diffusivity  $k = k_m / Pr$  with the Prandtl number  $Pr$  given by

$$Pr = \max(1, \min(10, 6.6 Ri)) , \quad Ri = N^2 \max\left((\partial_z \mathbf{u})^2, \epsilon_{iw} / k_m\right)^{-1} \quad (\text{A2})$$

This formulation for  $Pr$  and Richardson number  $Ri$  yields an equivalent interior mixing efficiency of 0.2. Note that in the reference experiments,  $\epsilon_{iw} = 0$  and we set  $E_{tke} = \max(E_{tke}, 10^{-6} \text{ m}^2/\text{s}^2)$  at each time step, since the production of  $E_{tke}$  by the shear of the mean flow is too low in the interior. The choice of the mixing length scale  $L$  follows Blanke and Delecluse (1993) (their Eqs. 2.27 to 2.30). The closure has been implemented together with IDEMIX in the Community Vertical Mixing Project (CVMix) (Griffies et al., 2015).

## Appendix B IDEMIX closure

IDEMIX (Internal Wave Dissipation, Energy and Mixing) is an internal wave model based on the radiative transfer equation, the spectral energy balance equation of internal gravity waves (Olbers & Eden, 2013). Several simplifications, most notably the integration in wavenumber space, reduce the complexity of the radiative transfer equation and lead to a partial differential equations for wave energy compartments that are simple enough to be solved online in global ocean general circulation models. Several different versions of IDEMIX have been proposed, including a low mode tidal and near-inertial wave compartment with explicitly resolved horizontal propagation (Eden & Olbers, 2014), a version including the effect of wave drag on the mean flow (Olbers & Eden, 2017; Eden & Olbers, 2017), and a version including a compartment for lee waves (Eden et al., 2021). In this study, however, we use the simplest IDEMIX approach (Olbers & Eden, 2013) as implemented in CVMix. It is given by

$$\partial_t E_{iw} = \partial_z (c_0 \tau_v \partial_z c_0 E_{iw}) + \nabla_h \cdot \tau_h v_0 \nabla_h v_0 E_{iw} - \epsilon_{iw} , \quad \epsilon_{iw} = \mu_0 f_e \frac{m_*^2}{N^2} E_{iw}^2, \quad (\text{B1})$$

where  $E_{iw}$  is a (single) wave energy compartment, and  $c_0$  and  $v_0$  bulk group velocities in vertical and horizontal direction, respectively, calculated assuming a certain spectral shape of the wave field, i.e. the Garrett-Munk (GM) model spectrum (Cairns & Williams, 1976; Munk, 1981).  $\epsilon_{iw}$  represents the dissipation of wave energy by wave breaking following (Heney et al., 1986) with  $f_e = |f| \text{acosh}(N/|f|)$ , and is also used for the so-called fine-structure parameterization (Kunze, 2017). Note that this form for  $\epsilon_{iw}$  was validated recently by Eden et al. (2019) by numerical evaluation of the scattering integral for wave-wave interactions.

The following parameters are contained in our simple IDEMIX closure:

- $\tau_v$  is a time scale on which wave-wave interactions lead to a symmetrization of the energy compartments of up- and downward propagating waves.
- $\tau_h$  is a corresponding time scale for eliminating lateral anisotropy.
- $\mu_0$  is related to the dissipation of internal wave energy by wave-wave interactions.

- $j_*$  is the equivalent mode number scale, related to the roll-off wavenumber  $m_*$  in the GM model spectrum by  $m_* = N/c$  with  $c = \int N/(j_*\pi)dz$ .

The parameter settings that lead to the best agreement with maps of wave energy and  $E_{tke}$  dissipation rates estimated from Argo float profiles are  $\tau_v = 2$  d,  $\tau_h = 15$  d,  $\mu_0 = 1/3$  and  $j_* = 5$  (Pollmann et al., 2017). Sensitivity tests in Pollmann et al. (2017) indicate that variations of  $\tau_v$  and  $\tau_h$  have very little impact on the average wave energy levels and TKE dissipation rates, whereas variations of  $j_*$  have the largest. Through its impact on the representative vertical group velocity, higher values of  $j_*$  will reduce the upper-ocean internal wave energy levels.

The generation of internal wave energy is accounted for in the vertical boundary conditions for the flux divergence term on the left-hand side of Eq. B1: at the surface, wind stress fluctuations create near-inertial oscillations of the mixed layer that can radiate internal waves of near-inertial frequency into the ocean interior, and at the bottom, the interaction of barotropic tidal currents with rough seafloor topography leads to the formation of internal tides. For the former, we update the maps used by Olbers and Eden (2013) and take instead the fraction of wind power input into near-inertial motions that leaves the mixed layer calculated by Rimac et al. (2013) and shown in Fig. 1(a). For the latter, we use three different maps, which are shown in Fig. 1(b-d).

## Appendix C Tidal forcing

Tidal forcing in IDEMIX is a two-dimensional map of the barotropic-to-baroclinic energy conversion applied at the bottom. This energy conversion can be estimated in several ways: from linear theory (Bell, 1975a, 1975b), from a simple scaling based on linear theory to describe the dissipation in barotropic tide models (Arbic et al., 2018), or from three-dimensional numerical simulations forced with the lunisolar tidal potential (Niwa & Hibiya, 2011; Müller et al., 2010; Buijsman et al., 2020).

Forcing A is a simple relation for the barotropic-to-baroclinic tidal energy conversion based on linear theory:

$$E_f \sim \frac{1}{2}\rho_0 k_{topo} h^2 N |\mathbf{u}|^2, \quad (\text{C1})$$

where  $h^2$  is the bottom roughness,  $\rho_0$  the density,  $N$  the buoyancy frequency,  $\mathbf{u} = (u, v)$  is the horizontal velocity vector and  $k_{topo}$  the topographic wavenumber treated as a free, spatially constant parameter (Jayne & St. Laurent, 2001). It was suggested by Jayne and St. Laurent (2001) to add an associated drag term  $-1/2k_{topo}h^2N\mathbf{u}$  as a sink to the barotropic shallow water momentum budget to account for the energy loss by internal tide generation, which led to a much better agreement with barotropic tide dissipation estimates obtained from satellite altimetry. The scaling Eq. C1 is often used in parameterizations of near-field tidal mixing in global numerical simulations (St. Laurent et al., 2002; Simmons et al., 2004; Griffies et al., 2015) and, evaluated globally for the Community Earth System Model (CESM) (Hurrell et al., 2013), also as tidal forcing in IDEMIX (Olbers & Eden, 2013). The latter is what we use as forcing A. As eq. C1 was obtained by neglecting any frequency dependence (Jayne & St. Laurent, 2001), forcing A represents all tidal constituents.

Forcing B is derived from linear theory, which builds on the work of Bell (1975a, 1975b). While Bell assumes an infinitely deep ocean, Llewellyn Smith and Young (2002) as well as Khatiwala (2003) considered a finitely deep ocean and derived the conversion into different vertical normal modes. These expressions or variants thereof have been evaluated globally a number of times: Nycander (2005), for example, performed global calculations for the 8 major constituents using Bell's theory, to which he applied a correction factor to mimic the behavior in a finitely deep ocean. Falahat et al. (2014) calculated the conversion globally for the first 10  $M_2$ -tide modes using the approach of Llewellyn Smith

707 and Young (2002). All linear theory approaches rely on several assumptions, i.a. that  
 708 the topography be subcritical, the topographic obstacles be much smaller than the wa-  
 709 ter depth, and the tidal excursion be small. To date, there is no analytically sound deriva-  
 710 tion of how to correct the relevant equations in cases when these assumptions are vio-  
 711 lated; instead, the calculations are performed everywhere and empirical corrections are  
 712 added later. The advantage of the linear theory approach is that topography input of  
 713 very high resolution can be used at reasonable computational costs. Here, we use the non-  
 714 modal linear theory estimates of Nycander (2005) as calculated by Falahat et al. (2014)  
 715 as forcing B, which represent the eight major tidal constituents  $M_2$ ,  $S_2$ ,  $N_2$ ,  $K_2$ ,  $K_1$ ,  $O_1$ ,  
 716  $P_1$ ,  $Q_1$ .

717 Forcing C is derived from a three-dimensional numerical model forced with the lun-  
 718 isolar tidal potential. The advantage of this approach (Niwa & Hibiya, 2011; Müller  
 719 et al., 2010; Buijsman et al., 2020) is that all the assumptions inherent in linear theory  
 720 are irrelevant, but on the downside, not all modes are resolved and other assumptions  
 721 to deal with the dissipation of the internal tide energy need to be invoked. For forcing  
 722 C, we consider the  $M_2$ -tide generation in the STORMTIDE2 simulation (Li & von Storch,  
 723 2020). STORMTIDE2 was performed using the primitive-equation model MPI-OM (Max-  
 724 Planck-Institute Ocean Model) (Marsland et al., 2003; Jungclaus et al., 2006) with a hor-  
 725 izontal resolution of  $0.1^\circ$  and 40 vertical levels, which resolves the lowest modes of the  
 726  $M_2$ -tide. Tides are excited by applying the full luni-solar tidal potential, parameteriz-  
 727 ing self-attraction and loading effects following Thomas et al. (2001). After a 33-year long  
 728 spin-up with a climatological forcing of daily resolution (Röske, 2006), the model is forced  
 729 by the 6-hourly NCEP/NCAD reanalysis-1 (Kalnay et al., 1996) and integrated for the  
 730 years 1981-2012. The barotropic-to-baroclinic energy conversion of the  $M_2$ -tide was eval-  
 731 uated for the final year of this period. Li et al. (2015) show that the relatively similar  
 732 STORMTIDE simulation fully resolves the propagation of the first two  $M_2$  tide modes.  
 733 It is likely that more modes are resolved when it comes to their generation, but it is un-  
 734 clear how many exactly. Because the lowest modes carry most of the energy, we will in  
 735 our comparison of the different tidal forcings for IDEMIX not make any correction for  
 736 the unresolved higher  $M_2$ -modes and only add the seven other constituents of the com-  
 737 putation by Nycander (2005) to obtain a total forcing agreeing with forcings A and B.

## 738 Open Research Section

739 The model code of ICON is available to individuals under licenses ([https://mpimet](https://mpimet.mpg.de/en/science/modeling-with-icon/code-availability)  
 740 [.mpg.de/en/science/modeling-with-icon/code-availability](https://mpimet.mpg.de/en/science/modeling-with-icon/code-availability)). By downloading the  
 741 ICON source code, the user accepts the licence agreement. The model code for FESOM  
 742 is available under: <https://zenodo.org/record/7737061>. The model code for MIT-  
 743 gcm can be found under <https://github.com/MITgcm/MITgcm>, specific modifications,  
 744 configuration, and plotting scripts can be found under [https://github.com/mjlosch/](https://github.com/mjlosch/MITgcm/tree/idemix_test_runs)  
 745 [MITgcm/tree/idemix\\_test\\_runs](https://github.com/mjlosch/MITgcm/tree/idemix_test_runs).

746 The source code of the specific ICON-O version used in this study, the configura-  
 747 tion files for the ICON simulations, and the post-processing scripts for ICON, FESOM  
 748 and the observational data can be found under [https://hdl.handle.net/21.11116/](https://hdl.handle.net/21.11116/0000-000C-DE5C-4)  
 749 [0000-000C-DE5C-4](https://hdl.handle.net/21.11116/0000-000C-DE5C-4). The ICON plots were made by making use of the ICON post-processing  
 750 toolbox pyicon (<https://gitlab.dkrz.de/m300602/pyicon>) and the FESOM plots were  
 751 made by making use of tripyview (<https://github.com/FESOM/tripyview>).

752 The CVMix implementation of IDEMIX and the TKE scheme which are used by  
 753 ICON and FESOM can be found within the corresponding model source codes and un-  
 754 der <https://github.com/nbruegge/CVMix-src>. MITgcm used an equivalent implemen-  
 755 tation of IDEMIX and the TEK scheme which can be found within the MITgcm source  
 756 code (see link above).

The tidal forcing of Nycander (2005) and Falahat et al. (2014) was obtained from <https://www.seaonoe.org/data/00470/58153/>, using the corrected form of the modal calculations of Falahat et al. (2014) provided by de Lavergne et al. (2019). The tidal forcing based on the scaling by Jayne (2009) is the same as used in CESM simulations, which we obtained from their subversion server <https://svn-ccsm-inputdata.cgd.ucar.edu/trunk/inputdata/ocn/pop/gx1v6/forcing/>.

The observational references were obtained from <https://ftp.nwra.com/outgoing/kunze/iwturb/> (Kunze, 2017). The global topography dataset of Becker et al. (2009) can be downloaded from [https://topex.ucsd.edu/marine\\_topo/](https://topex.ucsd.edu/marine_topo/).

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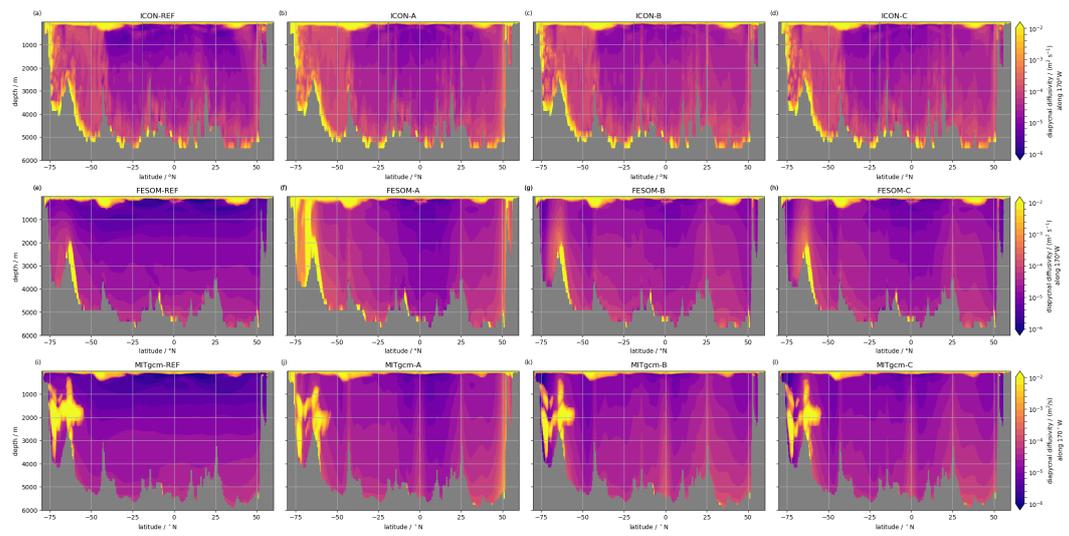
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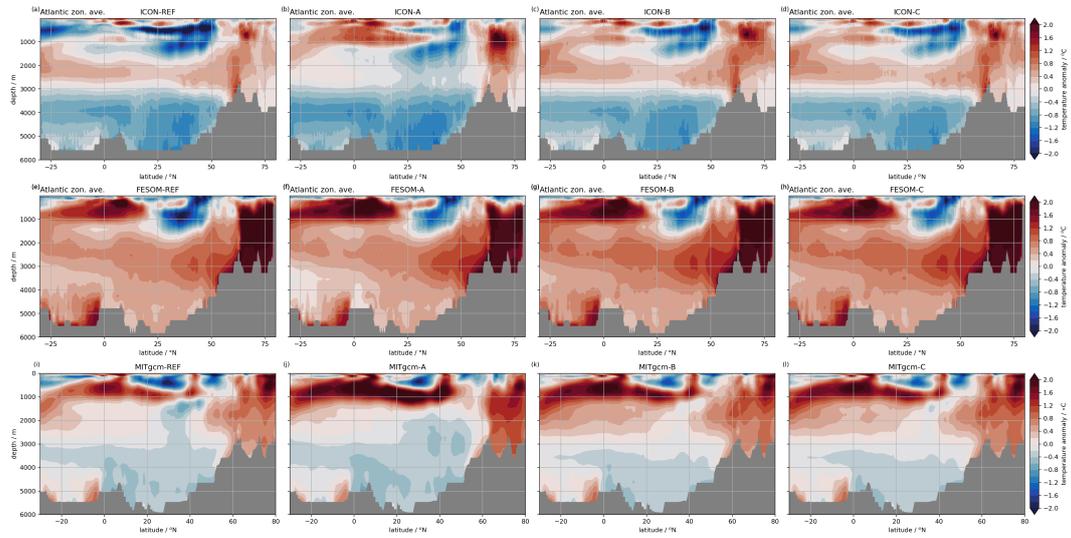
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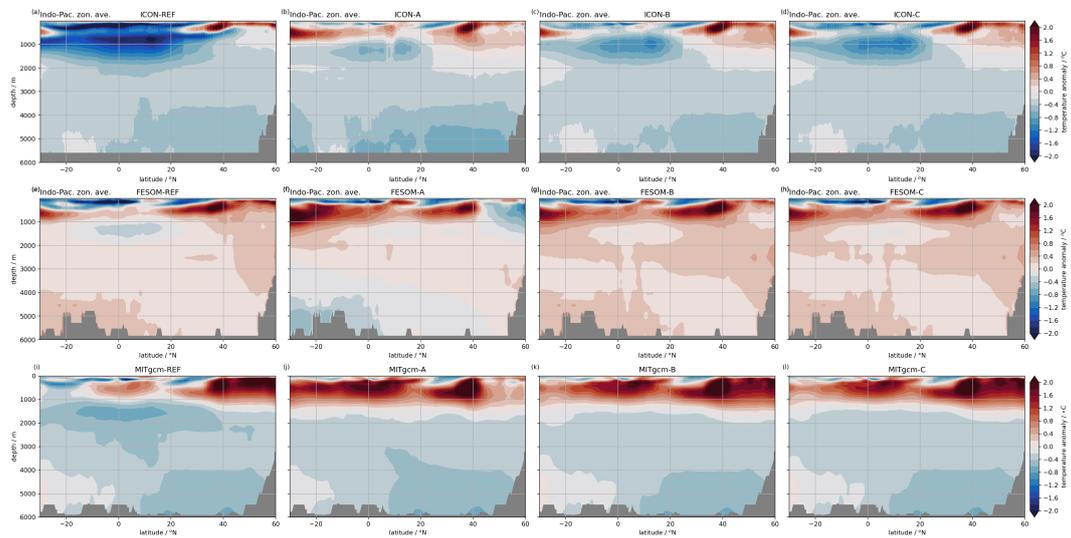
## Supplementary figures



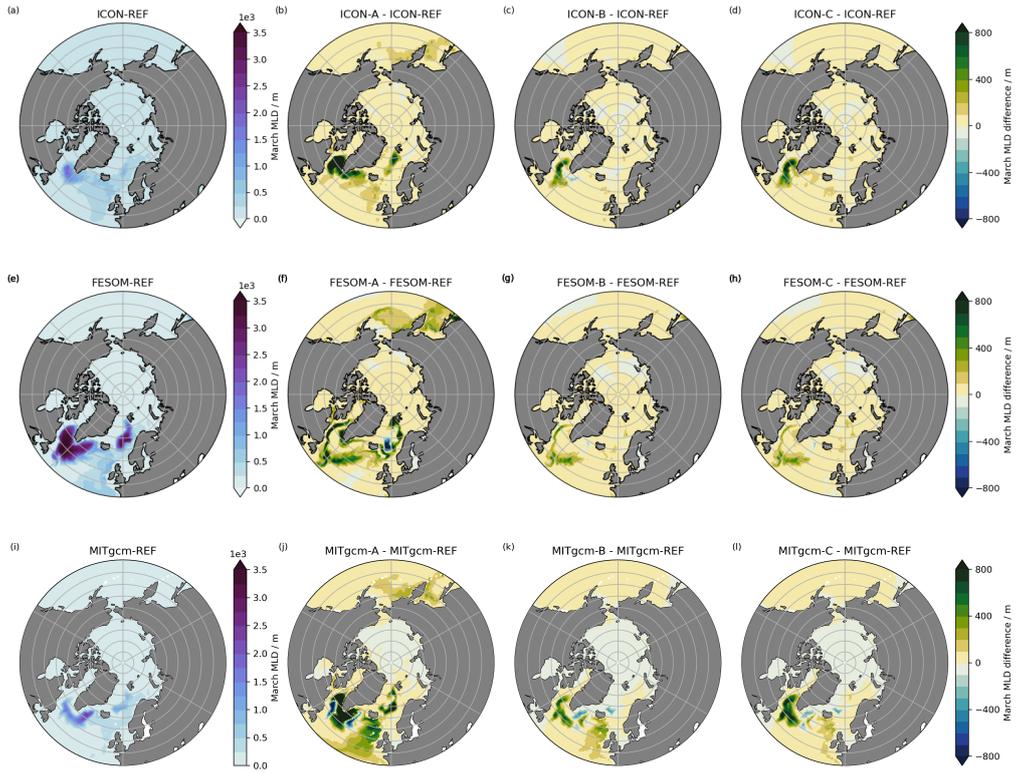
**Figure C1.** Diapycnal diffusivity  $\kappa$  along  $170^\circ \text{W}$  for ICON (a-d), FESOM (e-h) and MITgcm (i-l).



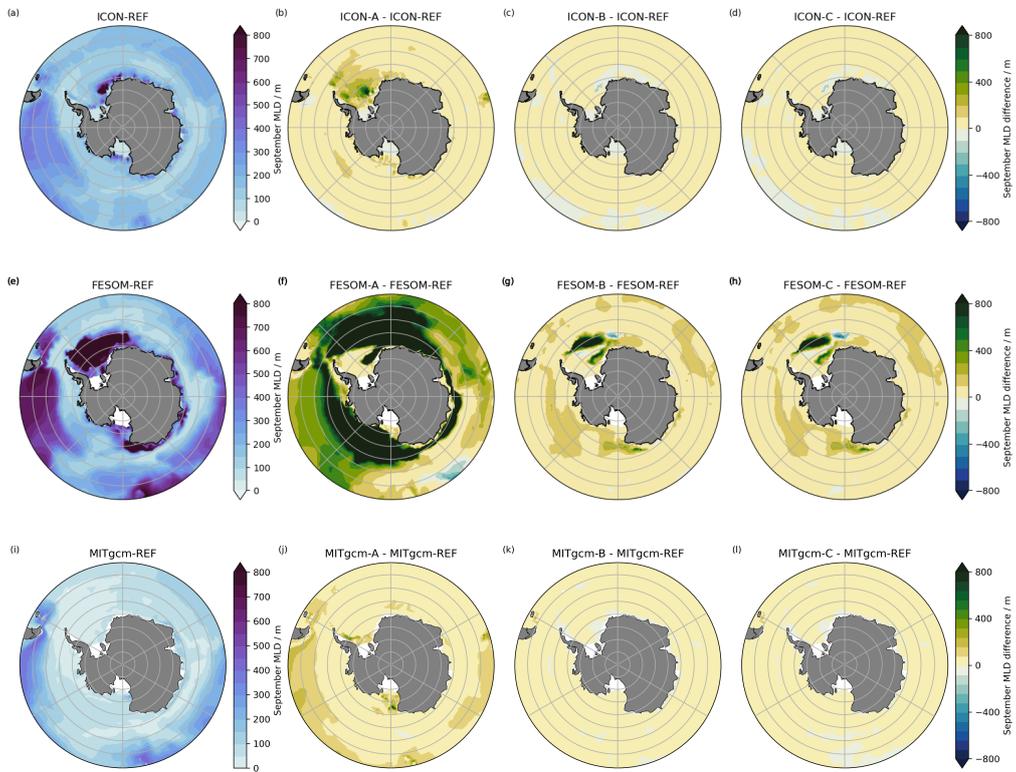
**Figure C2.** Atlantic zonal average of the temperature bias with respect to initial conditions for ICON (a-d), FESOM (e-h) and MITgcm (i-l).



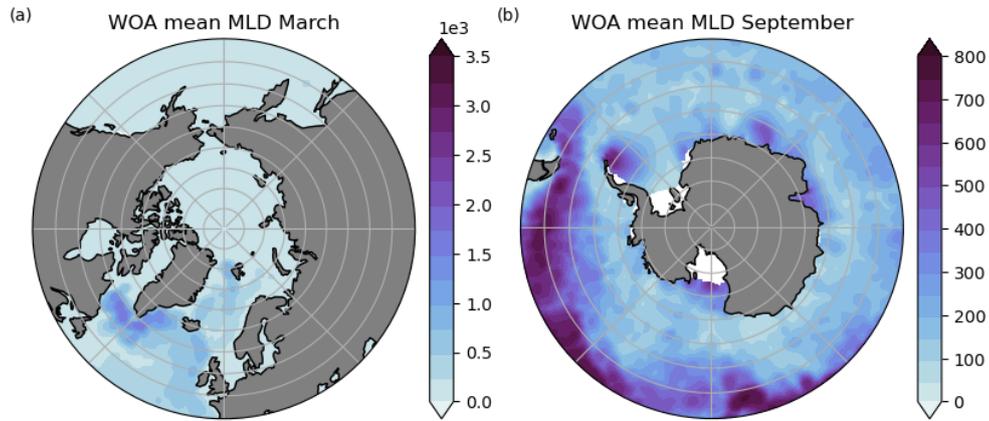
**Figure C3.** Indo-Pacific zonal average of the temperature bias with respect to initial conditions for ICON (a-d), FESOM (e-h) and MITgcm (i-l).



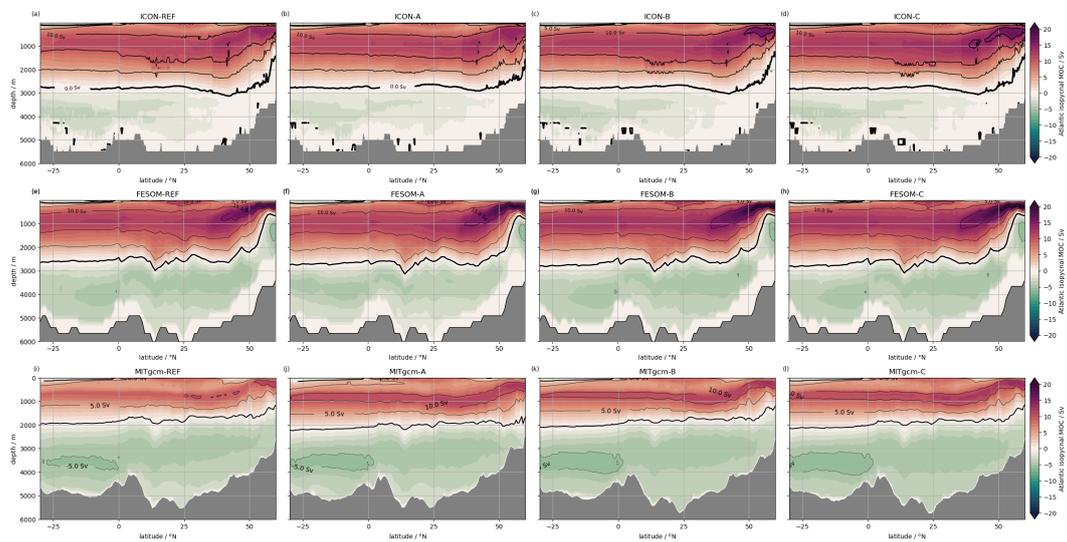
**Figure C4.** Mean mixed layer depth in March.



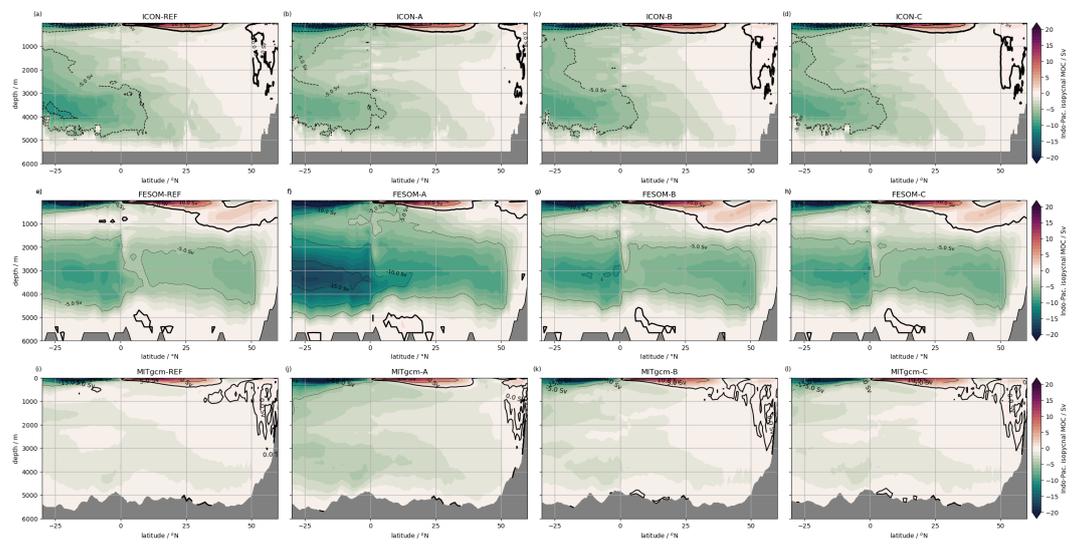
**Figure C5.** Mean mixed layer depth in September.



**Figure C6.** Mixed layer depth from the World Ocean Atlas 2018 (Locarnini et al., 2018; Zweng et al., 2019) for March (a) and September (b).



**Figure C7.** Atlantic meridional overturning in density space remapped to depth levels for ICON (a-d), FESOM (e-h) and MITgcm (i-l).



**Figure C8.** Indo-Pacific meridional overturning in density space remapped to depth levels for ICON (a-d), FESOM (e-h) and MITgcm (i-l).