

The Sensitivity of an Idealized Weddell Gyre to Horizontal Resolution

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1 **The Sensitivity of an Idealized Weddell Gyre to**
2 **Horizontal Resolution**

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6 **Key Points:**

- 7 • An idealized model of the Weddell Gyre demonstrates that the gyre transport can
8 be extremely sensitive to horizontal resolution.
- 9 • The gyre is strongest at eddy-permitting resolutions where the meridional den-
10 sity gradients are largest and the stratification is weakest.
- 11 • The depth-varying component of the Weddell Gyre is controlled by the density
12 structure and the bottom flow is sensitive to explicit eddies.

Abstract

Estimates of the Weddell Gyre transport vary widely between climate simulations. Here, we investigate if inter-model variability can originate from differences in the horizontal resolution of the ocean model. We run an idealized model of the Weddell Gyre at eddy-parameterized, eddy-permitting, and eddy-rich resolutions and find that the gyre is strongly sensitive to horizontal resolution. The gyre transport is largest at eddy-permitting resolutions (45 Sv with a noisy bathymetry) and smallest at eddy-parameterized resolutions (12 Sv). The eddy-permitting simulations have the largest horizontal density gradients and the weakest stratification over the gyre basin. The large horizontal density gradients induce a significant thermal wind transport and increase the mean available potential energy for mesoscale eddies. The distribution of eddy kinetic energy indicates that explicit eddies in simulations intensify the bottom circulation of the gyre via non-linear dynamics. If climate models adopt horizontal resolutions that the Weddell Gyre is most sensitive to, then simulations of the Weddell Gyre could become more disparate.

Plain Language Summary

The Weddell Gyre is a large horizontal circulation in the southern hemisphere. The gyre is exposed to low atmospheric temperatures and lies under extensive sea ice. Extremely dense water forms in the Weddell Sea, which the Weddell Gyre exports to the global ocean. These exported dense water masses change the Earth's climate by altering the total heat and carbon content in the global ocean. Between climate simulations, the volume of water transported by the Weddell Gyre varies significantly: we investigate if this variability can originate from differences in the horizontal spatial resolution of the ocean models. Using a simplified model of the Weddell Gyre, we find that the intensity of the circulation is extremely sensitive to the horizontal resolution. The circulation is particularly strong at intermediate resolutions, where only the largest ocean eddies are resolved. At intermediate resolutions, horizontal density gradients are the largest and the vertical density gradients are the smallest; this unique density structure allows for a particularly strong Weddell Gyre circulation. These results have important implications for long-range ocean climate projections.

1 Introduction

The Weddell Gyre is the largest subpolar gyre in the southern hemisphere and spans an area of approximately six million square kilometers in the Atlantic sector of the Southern Ocean (Vernet et al., 2019). Buoyancy forcing in this region is intense as atmospheric temperatures are low and sea ice formation is extensive. The sea ice insulates the ocean from the cold atmosphere but offshore winds can sustain coastal polynyas which enable strong buoyancy forcing through brine rejection (Haid & Timmermann, 2013; Abernathey et al., 2016). The Weddell Gyre also lies immediately south of the Antarctic Circumpolar Current (ACC), the strongest current in the global ocean.

Extremely dense water masses are produced in the Weddell Gyre as small bodies of water are exposed to intense buoyancy forcing for a prolonged period of time. Of particular interest is the production and export of Antarctic Bottom Water (AABW), which contributes to the southern closure of the global overturning circulation when exported northwards (Sloyan & Rintoul, 2001; Talley et al., 2003; Lumpkin & Speer, 2007; J. Marshall & Speer, 2012). The Weddell Gyre strength can control the variability of dense water export (Meijers et al., 2016) and could potentially influence global overturning. There is some debate about how much AABW is produced and exported by the Weddell Gyre and a review can be found in Vernet et al. (2019).

60 The surrounding coastline and local topographic features shape the Weddell Gyre,
 61 as seen in Figure 1a. The southern limb of the gyre follows the border of the Antarctic
 62 mainland and the the western limb is steered north by the Antarctic Peninsula. The east-
 63 ern boundary of the Weddell Gyre is a dynamic feature and it is uncertain whether any
 64 topographic feature limits it. Estimates of the eastern boundary location range from 30°E
 65 (Deacon, 1979) to as far as 70°E (Park et al., 2001). Within this longitudinal range there
 66 is an abundance of eddies that allow exchange between the gyre and ACC (Schröder &
 67 Fahrbach, 1999; Park et al., 2001; Ryan et al., 2016).

68 Two zonally-elongated ridges act as partial barriers between the ACC and Wed-
 69 dell Gyre: the South Scotia Ridge in the west and the North Weddell Ridge in the east
 70 (Vernet et al., 2019). These ridge systems are typically within 1500 to 2000 m of the sea
 71 surface. Submarine ridges block deep currents from crossing the ACC-gyre interface and
 72 play a major role in setting the stratification across the entire region (Orsi et al., 1993;
 73 Wilson et al., 2022). Figure 1b illustrates the potential density based on hydrographic
 74 sections of the Weddell Gyre and ACC. The contours of potential density in the Wed-
 75 dell Gyre are domed and there is a steep meridional density gradient above the subma-
 76 rine ridge (approx. 54°S in Figure 1b). As seen in Figure 1b, this is true if a reference
 77 pressure of 0 (σ_0), 1000 (σ_1), or 2000 (σ_2) dbar is used when calculating the potential
 78 density. As a result, only the densest components of the circumpolar flow are exposed
 79 to the intense buoyancy forcing found near the sea surface of the Weddell basin and on
 80 the continental shelf (Naveira Garabato et al., 2016).

81 Measurements of the Weddell Gyre transport are limited and vary widely. Gordon
 82 et al. (1981) use wind stress data and apply Sverdrup balance to estimate the Weddell
 83 Gyre transport as 76 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3\text{s}^{-1}$) while questioning the validity of Sverdrup
 84 dynamics. Moorings and ship data provide lower estimates of the transport, for exam-
 85 ple, 20-56 Sv from Fahrbach et al. (1991) and 30 Sv from Yaremchuk et al. (1998). In
 86 Reeve et al. (2019), the Weddell Gyre transport is estimated to be 32 ± 5 Sv using Argo
 87 data (Argo, 2020). Although recent Argo data have significantly increased the number
 88 of available observations, coverage is still fairly limited within the Weddell Gyre. Argo
 89 data has contributed approximately 28,000 hydrographic profiles (south of 60°S) over a
 90 time period of 14 years with no measurements taken below 2000 dbar. Reeve et al. (2019)
 91 use the thermal wind relation to estimate the geostrophic velocity in the upper 2000 dbar
 92 of the Weddell Gyre and extrapolate over depth to estimate the full volume transport
 93 while relying on ship-based observations to estimate the extrapolation error.

94 Climate models disagree on the strength and shape of the Weddell Gyre and lim-
 95 ited winter-time observations make it difficult to assess model accuracy in this region.
 96 Wang (2013) studies fourteen CMIP5 simulations with horizontal grid spacings of 1° or
 97 larger and finds that the time-averaged Weddell Gyre transport ranges from approximately
 98 10 to 80 Sv. This is troubling as Meijers et al. (2016) suggest that variability in the ex-
 99 port of dense Weddell Sea slope water is closely tied to wind-driven acceleration of the
 100 Weddell Gyre’s western boundary current. Inconsistent Weddell Gyre circulations be-
 101 tween climate models may lead to inconsistent descriptions of the global overturning cir-
 102 culation and consequently inconsistent global heat, carbon, and freshwater budgets. In
 103 the South Scotia Ridge, there are deep passages that are important AABW export path-
 104 ways. These pathways cannot be resolved in 1° models but will be present in future gen-
 105 erations of climate models with a finer horizontal resolution. It is therefore possible that
 106 future climate models will be more sensitive to the Weddell Gyre transport than they
 107 are now.

108 Long time integrations of numerical ocean models under different climate forcing
 109 scenarios are prohibitively expensive to run at mesoscale eddy-resolving resolutions, but
 110 high resolution simulations are becoming increasingly affordable. Hewitt et al. (2020)
 111 comment that the average horizontal resolution of the ocean has increased with each it-
 112 eration of CMIP and this corresponds to an approximate doubling of horizontal resolu-

113 tion every ten years (Fox-Kemper, 2018). The majority of centres participating in CMIP6
 114 parameterize the effect of unresolved eddies, but there are now several ‘eddy-permitting’
 115 models that at least partially resolve the mesoscale eddies (Hewitt et al., 2020), taking
 116 into account the small Rossby deformation radius at these high latitudes (Hallberg, 2013).

117 Significantly more observations are needed to assess how mesoscale eddies influence
 118 the Weddell Gyre transport in the real ocean and there are not enough eddy-permitting
 119 climate simulations in CMIP6 to accurately assess how explicit eddies influence the Wed-
 120 dell Gyre in climate projections. When studying the role of mesoscale eddies in the Wed-
 121 dell Gyre, a hierarchy of models with different resolutions can be a powerful tool. In Adcroft
 122 et al. (2019), the results of a 0.5° and a 0.25° ocean-ice model (GFDL OM4.0) are com-
 123 pared. These two models only differ in resolution and physical parameterization and pro-
 124 duce similar density structures in the Weddell Sea at 42°W . Neme et al. (2021) study
 125 the Weddell Gyre more directly in another hierarchy of ocean-ice models, ACCESS-OM2
 126 (Kiss et al., 2020), and find that the time-averaged Weddell Gyre transport increases from
 127 34 to 41 Sv when the horizontal grid spacing is reduced from 1° to 0.1° . Models such as
 128 those studied by Adcroft et al. (2019) and Neme et al. (2021) offer realism but are less
 129 configurable and more computationally expensive than idealized models that might demon-
 130 strate similar behaviour.

131 In the idealized and eddy-permitting simulations by Wilson et al. (2022), the in-
 132 troduction of a zonal submarine ridge intensifies the Weddell Gyre. Wilson et al. (2022)
 133 also comment that the ACC and Weddell Gyre primarily interact through transient ed-
 134 dles on the eastern boundary of the zonal ridge. In this article, we aim to investigate how
 135 mesoscale eddies can influence the Weddell Gyre and its interaction with the ACC. We
 136 do this by varying the horizontal resolution of an idealized model over a wide range of
 137 horizontal grid spacings including: eddy-parameterized scales (80 and 40 km), eddy-permitting
 138 scales (10 and 20 km), and eddy-rich scales (3 km). The Weddell Gyre is found to be
 139 extremely sensitive to horizontal resolution and is strongest at eddy-permitting resolu-
 140 tions.

141 The article is structured as follows. In Section 2 we describe the idealized model
 142 used in this study and in Section 3 we describe the three experiments that are carried
 143 out. In Section 4 we present our results including a thermal wind decomposition of the
 144 Weddell Gyre and ACC transport. In Section 5, we discuss how mesoscale eddies can
 145 strengthen the flow at the sea floor and the missing physics in our model design. Clos-
 146 ing remarks are made in Section 6.

147 2 Model design

148 The experiments presented in this article are performed in the NEMO Community
 149 Ocean model (Madec et al., 2019). NEMO has been used for several idealized gyre stud-
 150 ies (Lévy et al., 2010, 2015; Ruggiero et al., 2015; Perezhogin, 2019; Styles et al., 2022)
 151 and the presented configuration is similar to the model used by Wilson et al. (2022). The
 152 configuration features a zonally periodic channel and a southern continental shelf which
 153 resembles the neighbouring coastline for the Weddell Gyre (see Figure 2a). Two large
 154 landmasses are present on the western margins of the model, with an opening that crudely
 155 represents the Drake Passage. Additional topographic features include a submarine ridge
 156 which extends eastwards from the idealized Drake Passage and a meridional sill in the
 157 Drake Passage that blocks f/H contours and regulates the ACC transport (f is the Cori-
 158 olis parameter and H is the ocean depth). The parameters for these topographic features
 159 and all other relevant fixed parameters can be found in Table 1. Throughout this arti-
 160 cle, the x coordinate is the zonal displacement from the eastern boundary of the Drake
 161 Passage and the y coordinate is the meridional displacement from the southern bound-
 162 ary of the Drake Passage (see axes in Figure 2a).

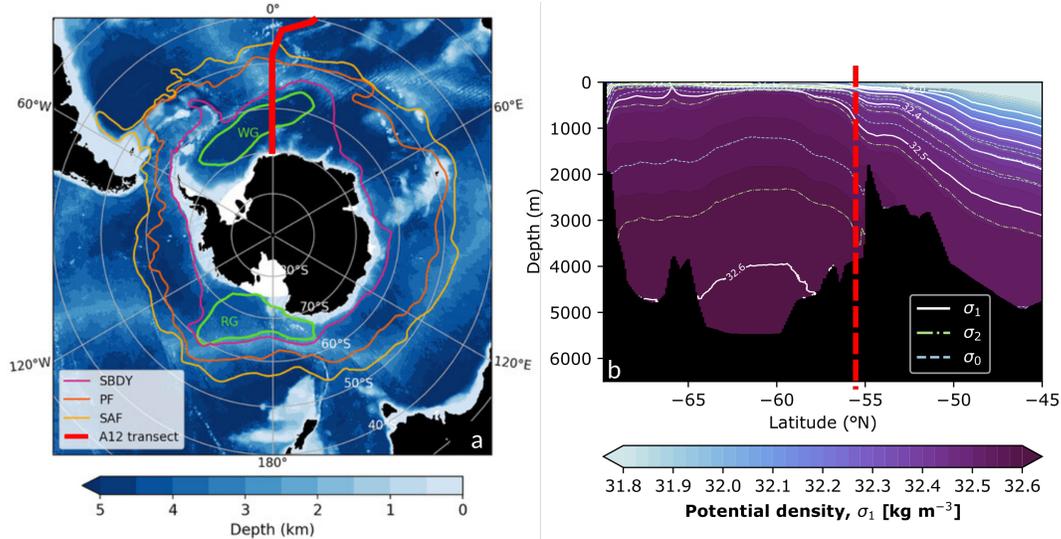


Figure 1. Bathymetric and hydrographic features of the Southern Ocean. (a) From Wilson et al. (2022), the bathymetry of the Southern Ocean. The contours mark three fronts of the ACC (Orsi et al., 1995): Southern Boundary (SBDY), Polar Front (PF), and Subantarctic Front (SAF). Outlines of the Weddell Gyre (WG) and Ross Gyre (RG) are also shown, using contours of satellite-based dynamic ocean topography (Armitage et al., 2018). (b) Potential density along the A12 transect [red line in (a)] calculated using conservative temperature and absolute salinity measurements from 10 hydrographic sections. The gridded hydrographic data originates from the GO-SHIP Easy Ocean product (Katsumata et al., 2022). White contours show potential density surfaces that are calculated with a reference pressure of 1000 dbar (σ_1). Green and blue contours show similar potential density surfaces that are calculated with a reference pressure of 2000 (σ_2) and 0 (σ_0) dbar respectively.

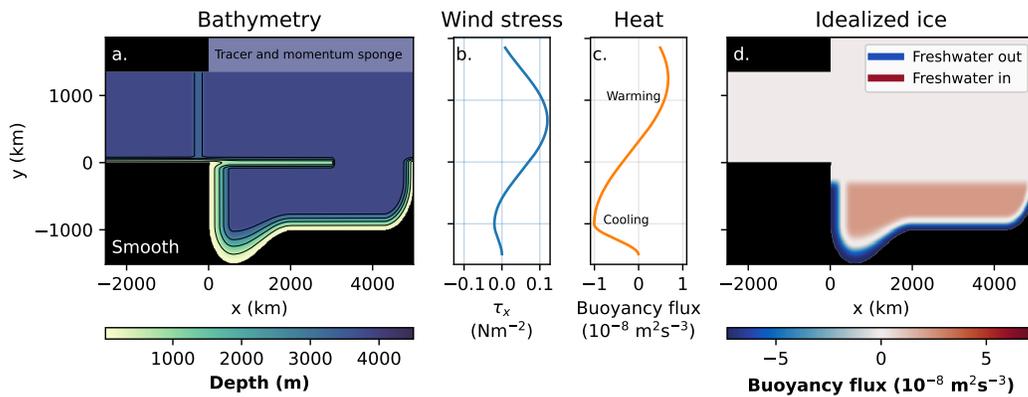


Figure 2. Summary of the model configuration. (a) Bathymetry of the model (without topographic noise) with contours at 1000 m intervals. (b) The zonal wind stress profile for the configuration. (c) The zonal heat flux profile for the configuration in units of buoyancy flux. (d) The freshwater fluxes used to represent sea ice for the configuration in units of buoyancy flux. In this model, a buoyancy flux of $10^{-8} \text{ m}^2\text{s}^{-3}$ corresponds to 14.9 Wm^{-2} of surface heating or approximately $4 \times 10^{-5} \text{ kg m}^{-2}\text{s}^{-1}$ of freshwater input (assuming a surface salinity of 35 psu).

Table 1. Summary of fixed parameters in the model. Δx is the horizontal grid spacing of the model in meters. (*) The sill height is varied for the ACC sensitivity experiment but is 500 m for all other experiments.

Model parameter	Value
Meridional domain size	3350 km
Zonal domain size	7520 km
Reference Coriolis parameter	$-1.3 \times 10^{-4} \text{ s}^{-1}$
Meridional gradient of Coriolis parameter	$9.6 \times 10^{-12} \text{ m}^{-1} \text{ s}^{-1}$
Momentum diffusivity (resolution dependent)	$(0.05 \Delta x) \text{ m}^2 \text{ s}^{-1}$
Tracer diffusivity (resolution dependent)	$(0.005 \Delta x) \text{ m}^2 \text{ s}^{-1}$
Maximum (smooth) ocean depth	4000 m
Number of model levels	31
Vertical resolution	10 - 315 m
Continental shelf width	300 - 600 km
Drake Passage zonal length	2520 km
Drake Passage meridional width	1350 km
Drake passage sill zonal width	500 km
Drake passage sill maximum height*	500 m
Submarine ridge zonal extent	3000 km
Submarine ridge meridional width	200 km
Submarine ridge maximum height	3000 m
Root mean square of topographic noise	100 m
Topographic noise length scales	240, 120, 60, 30, 9 km
Topographic noise relative amplitudes	8, 4, 2, 1, 0.3

163 The model has a regular horizontal grid with a horizontal grid spacing between 80
164 and 3 km, depending on the experiment (see Table 2). For context, on an ORCA grid
165 (Madec & Imbard, 1996) at 65°S the horizontal grid spacing is around 50 km in a 1° model,
166 12 km in a 0.25° model, and 5 km in a 0.1° model. All configurations use z -coordinates
167 and have 31 vertical model levels, with vertical spacing that is approximately 10 m near
168 the sea surface, 315 m near the sea floor, and partial cells are used to represent the vary-
169 ing sea floor. By only varying the horizontal resolution, we are only exploring dynam-
170 ics that do not require higher order vertical modes. The configuration exists on a beta
171 plane where the Coriolis parameter varies linearly with the meridional coordinate, y , around
172 its value at 65°S ($y=0$ in Figure 2). The model uses a free slip condition on lateral bound-
173 aries and applies a linear friction to the bottom boundary with a linear friction coeffi-
174 cient of $4 \times 10^{-4} \text{ m s}^{-1}$. A simplified linear equation of state is used with a thermal ex-
175 pansion coefficient of $a_0 = 2.8 \times 10^{-4} \text{ kg m}^{-3} \text{ K}^{-1}$ and a haline coefficient of $b_0 = 7.7 \times$
176 $10^{-4} \text{ kg m}^{-3} \text{ psu}^{-1}$. When using a linear equation of state, there is no distinction be-
177 tween conservative and potential temperature, nor is there a distinction between abso-
178 lute and practical salinity; therefore, in our results we will simply refer to temperature
179 and salinity. The horizontal diffusion of momentum and tracers is implemented with a
180 diffusivity that scales linearly with horizontal grid spacing (see Table 1). Subgrid scale
181 vertical mixing of momentum and tracers is represented by the Turbulent Kinetic En-
182 ergy scheme (Bougeault & Lacarrere, 1989; Gaspar et al., 1990).

183 The model is forced with a sinusoidal and zonal wind stress which only varies in
184 the meridional direction. The wind stress profile resembles the zonally and annually av-
185 eraged wind stress across the Southern Ocean (Figure 2b), with a maximum westerly wind
186 stress of 0.12 N m^{-2} over the center of the circumpolar channel and a peak easterly wind
187 stress of 0.02 N m^{-2} along the continental shelf. Similarly, the surface heat flux is also

188 sinusoidal and zonally uniform with a maximum surface warming of 10 Wm^{-2} at the north-
 189 ern boundary of the Drake Passage and a peak cooling of 15 W m^{-2} on the south con-
 190 tinental shelf. The surface heat flux is shown in units of buoyancy flux in Figure 2c.

191 The effect of sea ice on the salinity budget is simply represented by a surface fresh-
 192 water flux, as shown in units of buoyancy flux in Figure 2d. The freshwater fluxes re-
 193 semble the annual-average freshwater fluxes due to sea ice in the Southern Ocean, with
 194 net freshwater release in the Weddell basin and persistent sea ice formation on the south-
 195 ern continental shelf (Pellichero et al., 2018). Freshwater fluxes are the dominant buoy-
 196 ancy flux in the Weddell basin, as argued by Pellichero et al. (2018), but they do not ex-
 197 tend onto the submarine ridge in the idealized model. The domain area integral of fresh-
 198 water fluxes is identically zero to conserve the water content of the model. The model
 199 was also tested without the idealized sea ice. Removing the idealized sea ice does not
 200 change the gyre or ACC’s sensitivity to resolution but the isopycnals are less realistic
 201 above the continental shelf.

202 The northern margin of the model ($y > 1350 \text{ km}$) contains a sponge layer, which
 203 parameterizes the effect of the global ocean to the north. The horizontal flow is relaxed
 204 to rest, the salinity is relaxed to 35 psu at all depths, and the temperature is relaxed to
 205 the vertical profile,

$$206 \quad T(z) = T_{\text{top}} \exp(z/\delta_z), \quad (1)$$

207 where T is the temperature, z is the vertical coordinate, $T_{\text{top}} = 10^\circ\text{C}$ is the prescribed
 208 sea surface temperature and $\delta_z = 1500 \text{ m}$ is the decay length scale of the surface tem-
 209 perature. Consequently, the prescribed sea floor temperature is approximately 0°C . The
 210 momentum sponge has a maximum relaxation timescale of approximately 10 days and
 211 the tracer sponge has a maximum relaxation timescale of approximately 100 days. The
 212 sponge layer is 500 km wide and the relaxation timescale varies with y sinusoidally, in-
 213 creasing from zero to the maximum value.

214 The model’s initial state features a sinusoidal ACC with a peak zonal velocity of
 215 0.2 m s^{-1} in the channel center and a zero velocity at $y = 0 \text{ km}$ and $y = 1350 \text{ km}$. The
 216 ACC velocity profile decays with depth similarly to the temperature in the northern sponge
 217 (Equation 1). The initial state of the gyre basin (south of $y = 0$) is at rest. The ini-
 218 tial temperature profile is calculated using the thermal wind relation, integrating south
 219 from the northern sponge where the temperature profile matches Equation 1. Similarly
 220 to the sponge layer, the initial salinity is 35 psu everywhere in the domain.

221 In some experiments (see the next section) topographic noise is introduced to the
 222 bathymetry, as shown in Figure 3. The addition of weak topographic noise permits to-
 223 pographic interactions everywhere in the domain but only perturbs the larger scale bathy-
 224 metric features. The analytic noise field is generated using a zonally periodic and con-
 225 tinuous noise generation function, $\mathcal{N}(x, y)$, from OpenSimplex (Spencer, 2022). Noise
 226 is added at various length scales as shown below,

$$227 \quad \lambda(x, y) \propto \sum_{i=1}^{N_L} \mathcal{N}\left(\frac{x}{L_i}, \frac{y}{L_i}\right) L_i, \quad (2)$$

228 where λ is the final two dimensional noise function, L_i is the i^{th} length scale used (listed
 229 in Table 1), and N_L is the number of length scales used. The continuous function, $\lambda(x, y)$,
 230 is then evaluated on the grid used for each experiment and scaled so that the root mean
 231 square (rms) of the discrete noise field is 100 m in all configurations. Each length scale
 232 introduces a topographic gradient with a magnitude that is independent of L_i , as demon-
 233 strated below,

$$234 \quad \nabla \lambda(x, y) \propto \sum_{i=1}^{N_L} \nabla \mathcal{N}\left(\frac{x}{L_i}, \frac{y}{L_i}\right). \quad (3)$$

235 As seen in Figure 3, the maximum displacement caused by the noise field is approximately
 236 300 m and the structure of the continental shelf and other large topographic features is

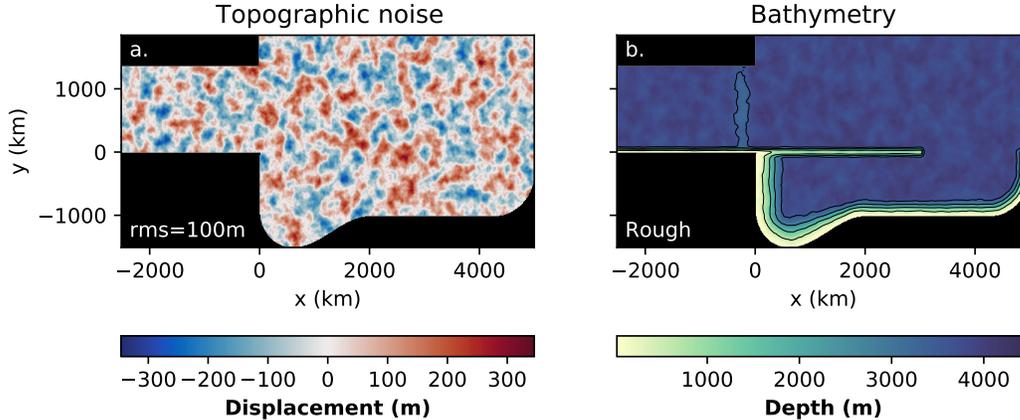


Figure 3. (a) Noise profile and (b) bathymetry for the Rough simulation with a horizontal grid spacing of 3 km. All discrete noise fields used in this article have a root mean square of 100 m and are based on the same continuous noise function, $\mathcal{N}(x, y)$.

237 not lost to the noise. In cases where the added noise would create islands in the domain,
 238 the noise is locally reduced to keep all topographic features submerged.

239 In Section 5, we will discuss the important differences between this idealized con-
 240 figuration and the real ocean and assess how the discrepancies may modify the results
 241 presented in this article.

242 3 Experimental setup

243 The model described in the previous section is computationally affordable and a
 244 wide parameter space can be explored. In total, 78 simulations were conducted with a
 245 minimum run time of 220 years. A summary of the experiments is shown in Table 2 and
 246 the computational cost is summarized in Appendix A. There are three sets of experi-
 247 ments: Smooth, Rough, and ACC sensitivity. The Smooth experimental series uses the
 248 bathymetry shown in Figure 2a and does not feature any topographic noise. The hori-
 249 zontal grid spacing is varied from 80 to 10 km and only the 80 and 40 km simulations
 250 include the Gent and McWilliams (1990) eddy parameterization (GM hereafter). The
 251 80 and 40 km simulations use a constant GM diffusivity parameter of $2000 \text{ m}^2 \text{ s}^{-1}$. The
 252 Rough series is exactly the same as the Smooth series but uses topographic noise, as shown
 253 in Figure 3, and the horizontal grid spacing varies from 80 to 3 km. The 3 km simula-
 254 tion is eddy-rich and computationally expensive so only one eddy-rich time integration
 255 could be completed. In Section 4.5, we also test various applications of GM on eddy-permitting
 256 and eddy-parameterized simulations to see how this affects our main results.

257 In this model the ACC is driven by wind-stress, surface buoyancy forces, and buoy-
 258 ancy forcing on the northern boundary. The ACC transport is not prescribed, so the ACC
 259 strength is free to respond to changes in the horizontal resolution. The ACC sensitiv-
 260 ity experiment series is designed to assess how strongly the Weddell Gyre and ACC trans-
 261 ports are coupled. Not only is a study of the gyre-ACC coupling scientifically interest-
 262 ing; it is also necessary to assess if the changes in the Weddell Gyre transport with res-
 263 olution are influenced by changes in the idealized ACC strength. In the ACC sensitiv-
 264 ity experiments, the height of the Drake Passage sill is varied from 500 m to 2500 m in
 265 intervals of 200 m: this modifies the strength of the simulated ACC in a way that does
 266 not modify the immediate conditions for the Weddell Gyre. For example, we cannot mod-

Table 2. Summary of the numerical experiments. Experiments marked with GM use the Gent and McWilliams (1990) eddy parameterization. In total 44 ACC sensitivity experiments were conducted (11 for each horizontal grid spacing between 80 and 10 km).

Experiment series	Horizontal grid spacing (km)					Topographic noise
	80 ^{GM}	40 ^{GM}	20	10	3	
Smooth	✓	✓	✓	✓	✗	✗
Rough	✓	✓	✓	✓	✓	✓
ACC sensitivity (11 variations of sill height)	✓	✓	✓	✓	✗	✓

267 ify the wind stress to change the ACC strength as this will alter the wind stress curl above
 268 the gyre and change the gyre strength directly. Modifying the wind stress would also be
 269 further complicated by eddy saturation effects in simulations with explicit eddies (D. P. Mar-
 270 shall et al., 2017).

271 4 Results

272 As seen in Figure 4, 200 years is a sufficient spin up time to assume a statistically
 273 steady Weddell Gyre and ACC transport. The transports in Figure 4 are calculated from
 274 annual averages of the depth-integrated stream function. The ACC transport is the zonal
 275 average of the stream function along the northern boundary of the domain and the Wed-
 276 dell Gyre transport is the maximum value of the stream function in the gyre basin (south
 277 of $y=0$). Only the gyre transport in the 3 km simulation shows a slight downward trend
 278 that does not alter the interpretation of the presented results. We also find that the spa-
 279 tial average of temperature and salinity are also statistically steady. Figure 5 shows the
 280 evolution of the domain-averaged temperature and salinity for configurations with a rough
 281 bathymetry. All configurations have statistically-steady temperature and salinity but the
 282 eddy-permitting configurations are on average 0.2-0.4 °C cooler than the eddy-parameterized
 283 and eddy-rich simulations.

284 All results presented in this section are time-averages taken from the final 20 years
 285 of each model run. As indicated by Figure 6, the 3 km simulation resolves a rich eddy
 286 field in the ACC and near the eastern boundary of the zonal ridge. Similar to Wilson
 287 et al. (2022), a weaker but qualitatively similar eddy field is partially resolved in 10 and
 288 20 km simulations. The distribution of eddy kinetic energy is discussed further in Sec-
 289 tion 5.1.

290 4.1 Transport sensitivity to resolution

291 In the Rough and Smooth configurations, the Weddell Gyre and ACC are strongly
 292 sensitive to resolution, as shown in Figure 7. The transports are calculated from the twenty
 293 year time-average of the depth-integrated stream function. Figure 7a shows how the Wed-
 294 dell Gyre transport increases as the resolution is doubled over smooth bathymetry. The
 295 time-averaged Weddell Gyre transport is 28.9 Sv in the 80 km simulation and increases
 296 to 54.7 Sv in the 10 km simulation. Introducing a rough bathymetry reduces all gyre trans-
 297 ports (17 Sv reduction for the 80 km case, 10 Sv reduction for the 10 km case) but also
 298 increases the Weddell Gyre’s sensitivity to resolution (Figure 7c). With a rough bathymetry,
 299 the time-averaged Weddell Gyre transport is 11.9 Sv in the 80 km configuration and then
 300 rapidly increases to 44.8 Sv in the 10 km configuration. For the Rough configuration,
 301 we have access to an eddy-rich simulation where the Weddell Gyre transport is 38.6 Sv.
 302 In this case, the transition from an eddy-permitting to an eddy-rich simulation reduces
 303 the Weddell Gyre transport by 6.2 Sv. The red error bars in Figure 7a through to 7d

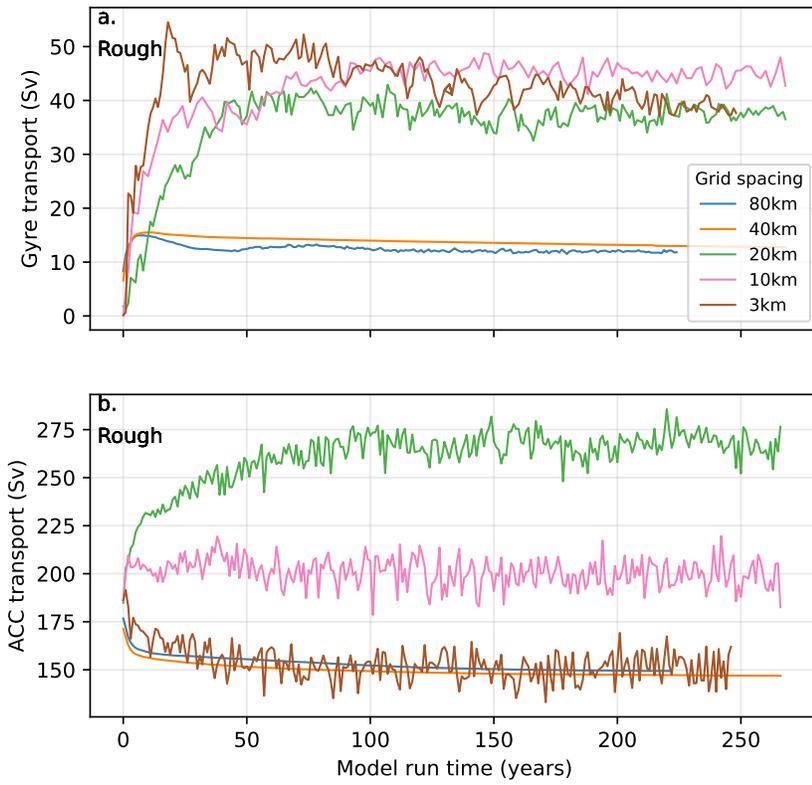


Figure 4. Evolution of the (a) Weddell Gyre and (b) ACC transports for configurations with rough bathymetry.

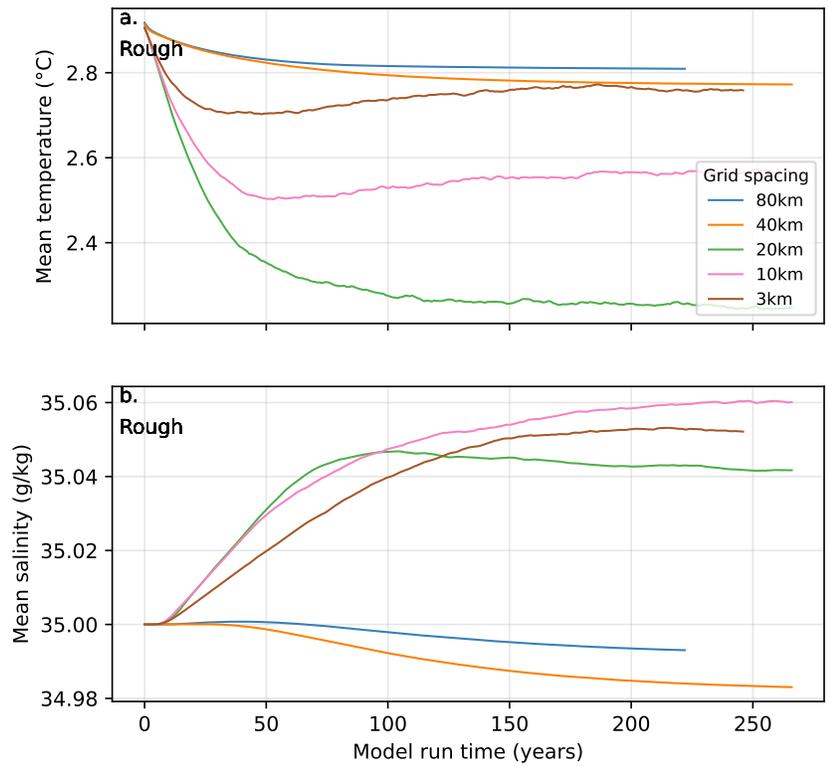


Figure 5. Evolution of the domain-averaged (a) temperature and (b) salinity for configurations with rough bathymetry.

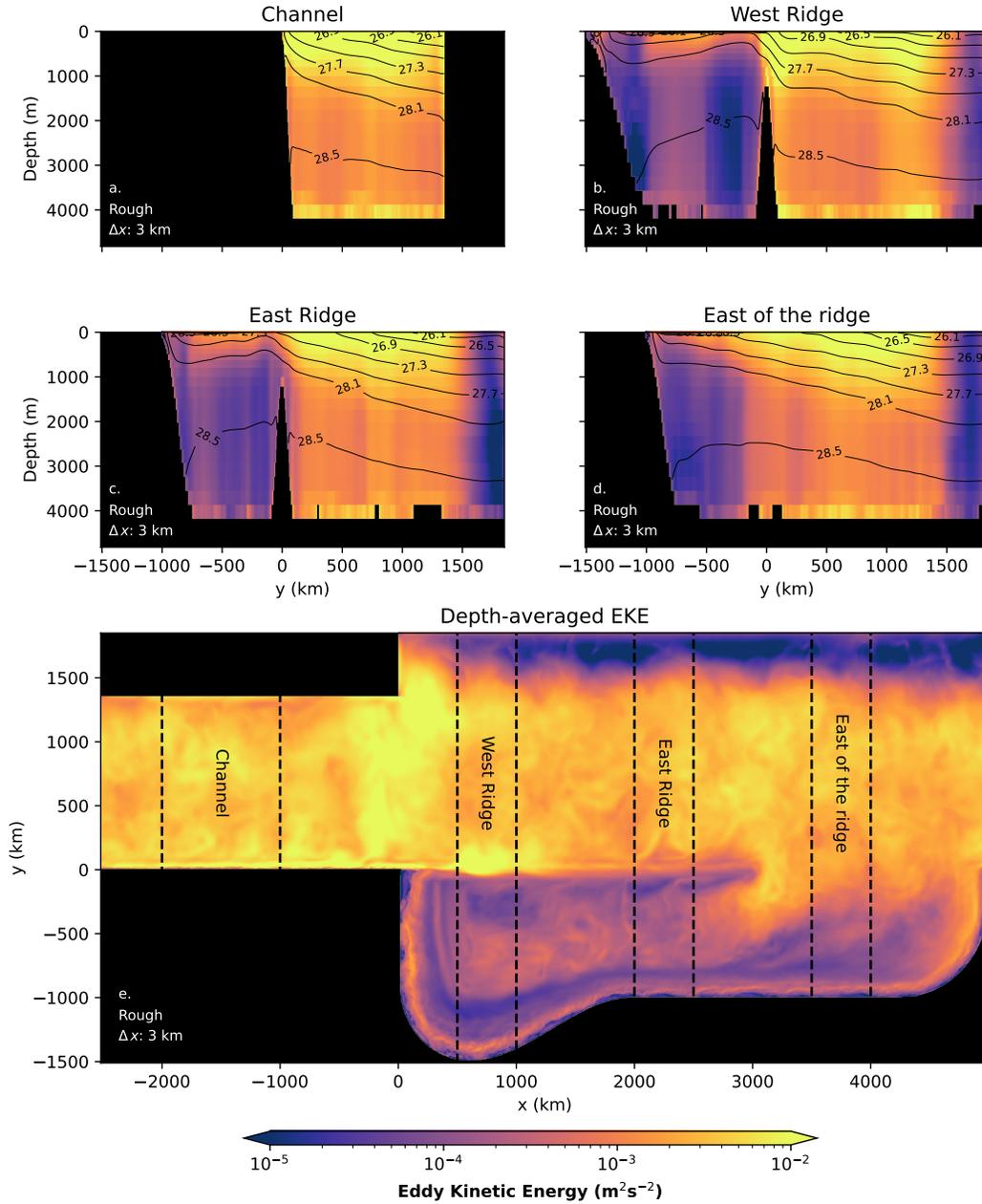


Figure 6. (a)-(d) Zonal averages and the (e) depth average of the eddy kinetic energy in a simulation with a 3 km horizontal grid spacing and rough bathymetry. The zonal averages in (a)-(d) are taken over various meridional sections that are shown in (e). The black contours in (a)-(d) show the zonally-averaged density surfaces across the meridional section.

304 show the interannual variability of the gyre and ACC transports during the experimen-
 305 tal period (two standard deviations either side of the average). The limited overlap of
 306 adjacent error bars suggests that the gyre and ACC transports for almost any year in
 307 our experiment are similar to the time-averaged results.

308 In Figure 7, the time-averaged stream function for the gyre shrinks and follows the
 309 bathymetry more closely when the resolution increases from eddy-parameterized (80 km)
 310 to eddy-permitting (10 km) scales. The boundary current that forms on the submarine
 311 ridge becomes particularly narrow and intense. In Section 4.3 we use a thermal wind de-
 312 composition to relate the transports and stream functions shown in Figure 7 to horizon-
 313 tal density gradients and the velocity at the sea floor.

314 The structure of the ACC is sensitive to the coarseness of the bathymetry. In the
 315 Smooth configurations, the ACC is deflected northwards by the ridge in the Drake Pas-
 316 sage but quickly returns to a zonal flow which results in large positive and negative merid-
 317 ional velocities in the ACC (Figure 7g). In the Rough configurations, the northward de-
 318 flection of the ACC is similarly severe but the topographic noise appears to dampen the
 319 ACC's return to a zonal flow (Figure 7h and 7i). The ACC's behaviour in the Rough
 320 configurations more closely resembles the ACC's abrupt equatorward deflection east of
 321 the Drake Passage and the current's more steady poleward trajectory while circumnav-
 322 igating Antarctica (see Figure 7 of Mazloff et al. (2010)).

323 The time-averaged ACC transport is also strongly sensitive to resolution in both
 324 Smooth and Rough configurations. In particular, the transition from a smooth to rough
 325 bathymetry intensifies the ACC transport at eddy-permitting resolutions (increasing from
 326 221.7 to 266.2 Sv at 20 km). In Section 4.3, the ACC transport will also be related to
 327 horizontal density gradients and the bottom velocity. With a maximum ACC transport
 328 of 266.2 Sv (20 km horizontal grid spacing, rough bathymetry) and a minimum of 147.7
 329 Sv (40 km horizontal grid spacing, rough bathymetry), it is important to assess if such
 330 large variations in the ACC transport modify the Weddell Gyre transport directly.

331 4.2 ACC sensitivity results

332 In the ACC sensitivity experiments, the strength of the ACC is varied by modi-
 333 fying the height of the sill in the idealized Drake Passage. As shown in Table 2, these
 334 experiments use horizontal grid spacings between 80 and 10 km and all experiments have
 335 a rough bathymetry.

336 The results of the ACC sensitivity experiment are shown in Figure 8 where we can
 337 immediately see that the gyre transport barely responds (~ 1 Sv) to large changes in
 338 the idealized ACC transport (~ 100 Sv) at all tested resolutions. The large changes in
 339 the Weddell Gyre transport with resolution are controlled by the gyre's direct sensitiv-
 340 ity to resolution and not the gyre's sensitivity to the ACC strength.

341 4.3 Thermal wind and bottom flow decomposition

342 In Section 4.1 we observe that the idealized Weddell Gyre and ACC transports are
 343 sensitive to horizontal resolution and are particularly strong at eddy-permitting resolu-
 344 tions. In this section, we relate the observed transports to the isopycnal structure of the
 345 circulation and the strength of the circulation at the sea floor. The depth-integrated ve-
 346 locity field is separated into depth-dependent and depth-independent components us-
 347 ing integration by parts,

$$348 \quad \mathbf{U} = \int_{-H}^{\eta} \mathbf{u} dz = \mathbf{u}_b H + \mathbf{u}_t \eta - \int_{-H}^{\eta} \frac{\partial \mathbf{u}}{\partial z} z dz, \quad (4)$$

349 where \mathbf{U} is the depth-integrated velocity field, η is the free surface height, \mathbf{u}_b is the ve-
 350 locity at the sea floor, and \mathbf{u}_t is the velocity at the free surface. We then use the follow-

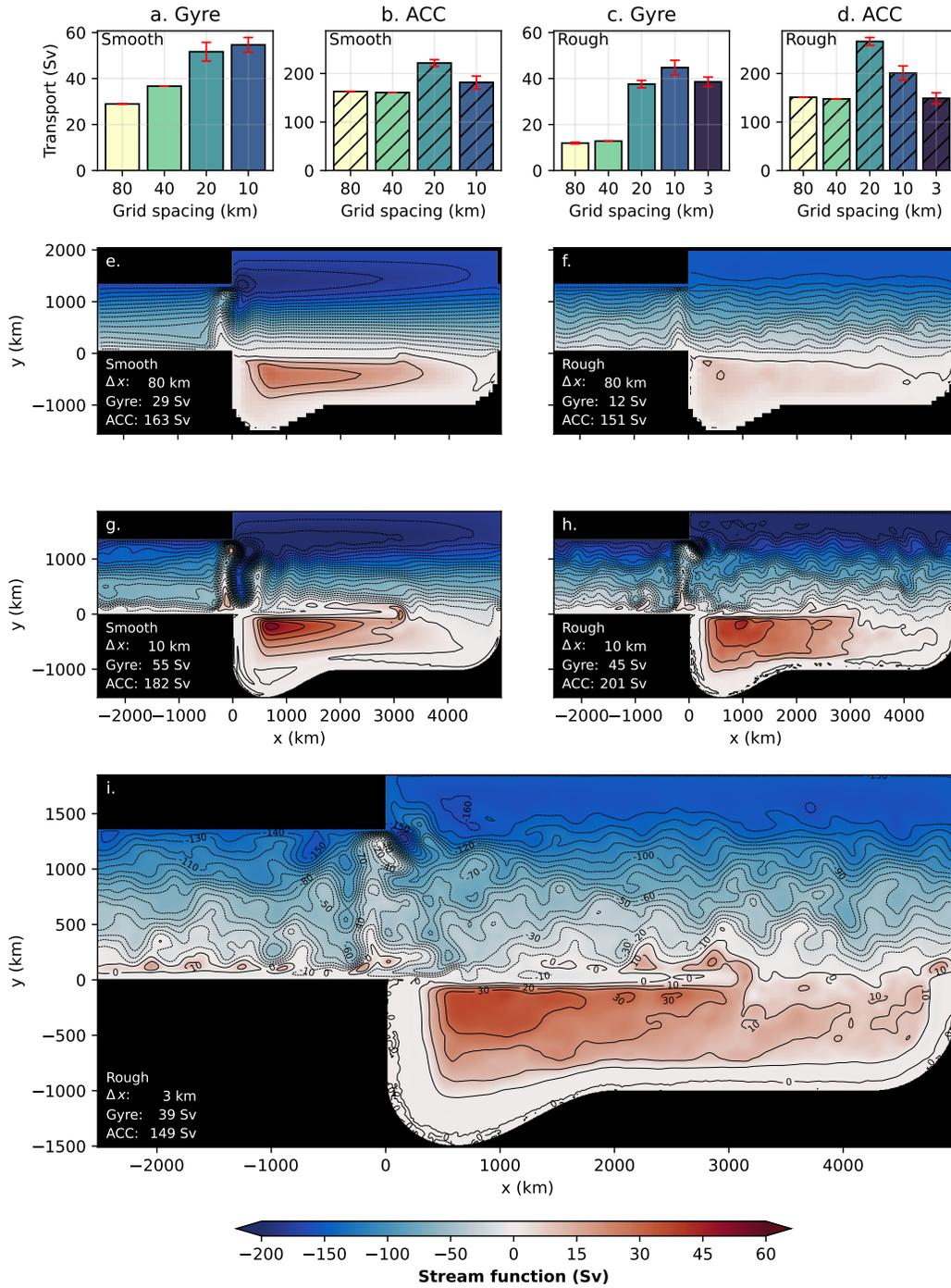


Figure 7. Sensitivity of the Weddell Gyre and ACC transport to resolution. (a)-(d) show the time-averaged gyre and ACC transport over smooth and rough bathymetry. Red error bars show the inter-annual variability of the transport (two standard deviations). (e)-(i) are the time-averaged stream functions from configurations at an eddy-parameterized resolution (80 km), an eddy-permitting resolution (10 km), and an eddy-rich resolution (3 km). The black contours have an interval of 10 Sv.

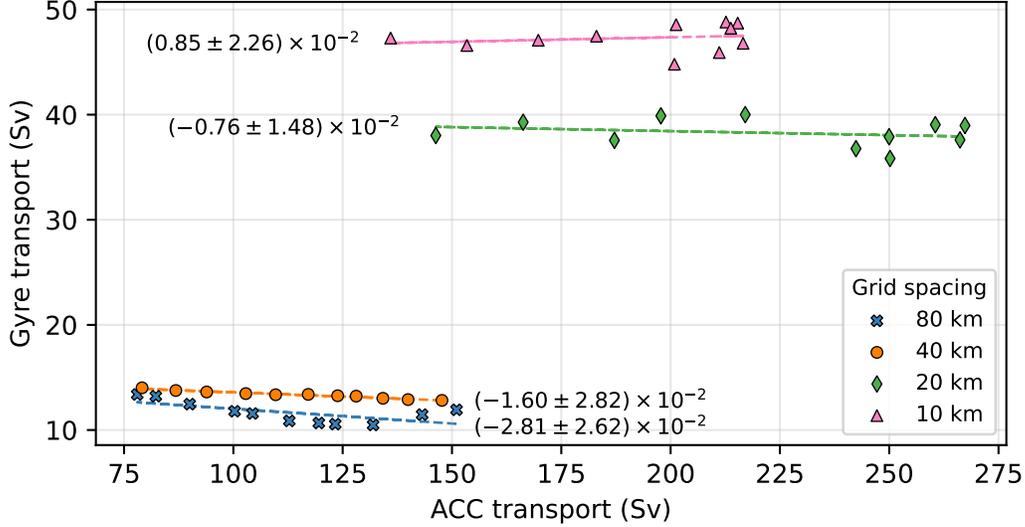


Figure 8. How the Weddell Gyre (WG) transport varies with respect to the ACC transport at several resolutions. The dashed line shows the straight line of best fit which is calculated using a least squares method. The value of the slope (dimensionless) and error is provided for each line. The gyre transport is insensitive to large variations in the ACC transport.

ing equation to describe how the the velocity field varies with depth,

$$f \frac{\partial \mathbf{u}}{\partial z} = -\frac{g}{\rho_0} (\hat{\mathbf{k}} \times \nabla_h \rho) + \mathcal{E}, \quad (5)$$

where g is the acceleration due to gravity, ρ_0 is the reference density, ∇_h is the horizontal gradient operator, ρ is the density, and \mathcal{E} is a residual function. Equation 5 is the thermal wind equation where all non-geostrophic terms, non-hydrostatic terms, and numerical errors are aggregated in \mathcal{E} . The model used in this article assumes hydrostatic balance so \mathcal{E} is free of non-hydrostatic terms.

By combining Equations 4 and 5 we can derive a full decomposition of the depth-integrated flow,

$$\mathbf{U} = \underbrace{\mathbf{u}_b H}_{\text{Bottom velocity}} + \underbrace{\frac{g}{\rho_0 f} \int_{-H}^{\eta} (\hat{\mathbf{k}} \times \nabla_h \rho) z dz}_{\text{Thermal wind}} + \mathbf{u}_t \eta + \mathcal{E}^z, \quad (6)$$

where \mathcal{E}^z is the depth-integrated and rescaled residual that still contains non-geostrophic terms and errors from the discretization. The free surface term, $\mathbf{u}_t \eta$, is negligible in all of the presented results.

Using Equation 6, we decompose the Weddell Gyre and ACC transport into depth-independent ($\mathbf{u}_b H$) and depth-dependent components (thermal wind and residual). The decomposed transports and the associated stream functions are shown in Figures 9 and 10 respectively. In order to calculate valid stream functions, a Helmholtz decomposition of each term in Equation 6 is determined using an elliptical solver and the divergent part is removed. The full circulation has a weak divergent component and when this component is removed the largest change in the stream function is 0.03 Sv. The full circulation is approximately incompressible, so only the incompressible components of the terms in Equation 6 influence the full circulation. Each component of the gyre transport shown

373 in Figure 9 is equal to the component’s stream function evaluated where the gyre’s total
 374 transport is calculated (black crosses, Figure 10). This definition was chosen as it de-
 375 scribes the decomposition of the quantity of interest, the gyre’s transport. As shown in
 376 Figure 10, the maximum of a specific component’s stream function in the gyre basin (blue
 377 crosses) does not necessarily align with the gyre centre. The ACC transport components
 378 shown are the zonal averages, but zonal variations of the decomposition are small and
 379 do not alter our interpretation of the results.

380 In all cases, the combined transport from the bottom flow and thermal wind com-
 381 ponent closely describes the total transport of the gyre (black crosses in Figure 9). This
 382 suggests that the residual terms are minor when considering the gyre and ACC trans-
 383 ports. The circulation associated with the residual term, \mathcal{E}^z , is weak across most of the
 384 horizontal domain meaning that the depth-integrated circulation can be described to lead-
 385 ing order using geostrophic assumptions. It is important to note that a small value of
 386 \mathcal{E}^z does not guarantee geostrophy at all depths. The residual is largest at lower resolu-
 387 tions, this may be caused by small departures from geostrophy through viscous effects
 388 or a larger numerical error that comes with a coarser grid.

389 Looking at the gyre transports, the relative significance of the bottom velocity and
 390 the thermal wind component depends on the coarseness of the bathymetry. In the Smooth
 391 configurations, the bottom velocity plays a dominant role in controlling the gyre trans-
 392 port and increases with horizontal resolution. In configurations with a rough bathymetry,
 393 the gyre transport from the bottom flow is reduced and comparable to the thermal wind
 394 component, but still increases with resolution. When a rough bathymetry is used, the
 395 thermal wind component of the gyre is particularly strong with a 10 km horizontal grid
 396 spacing and consequently the total gyre transport is particularly strong at eddy-permitting
 397 resolutions.

398 The decomposition of the ACC transport is also dependent on the coarseness of
 399 the bathymetry. In simulations with a smooth bathymetry, contributions to the ACC
 400 transport from the bottom flow and thermal wind components are similar in size (Fig-
 401 ure 9b). The bottom flow component shrinks with resolution and the thermal wind com-
 402 ponent is largest at an eddy-permitting resolution (20 km in Figure 9b). When a rough
 403 bathymetry is used, the ACC transport is almost entirely determined by the thermal wind
 404 component, which is even larger at eddy-permitting resolutions (see Figure 9d). The de-
 405 composition of the ACC with rough bathymetry is more realistic as observations (Chidichimo
 406 et al., 2014; Koenig et al., 2014; Donohue et al., 2016) suggest that the ACC transport
 407 is largely determined by a thermal wind (or often called ‘baroclinic’) component.

408 The shape of the stream function from the bottom flow and the thermal wind com-
 409 ponents differ. The thermal wind stream function (left column of Figure 10) features a
 410 gyre that lies over the basin interior and submarine ridge and is not west-intensified. Over
 411 the continental shelf there is a density-driven slope current that reverses direction south
 412 of the submarine ridge, which is a consequence of the idealized model design. An accu-
 413 rate slope current may require a wind stress that follows the continental shelf (Thompson
 414 et al., 2018). The stream function for the bottom flow transport (right column of Fig-
 415 ure 10) features a gyre that follows the bathymetry closely and is west-intensified. The
 416 submarine ridge blocks the deep current but the bottom flow is free to extend northwards
 417 into the ACC channel once it is far enough east.

418 Over the submarine ridge the thermal wind and bottom velocity stream functions
 419 reinforce each other, resulting in a particularly strong western and northern boundary
 420 current. In contrast the thermal wind and bottom velocity stream function are opposite-
 421 signed on the continental shelf, which limits the gyre’s presence over the continental shelf
 422 in all simulations. In higher resolution simulations, the bottom velocity stream functions
 423 uniquely feature intense recirculations to the east of the Drake Passage.

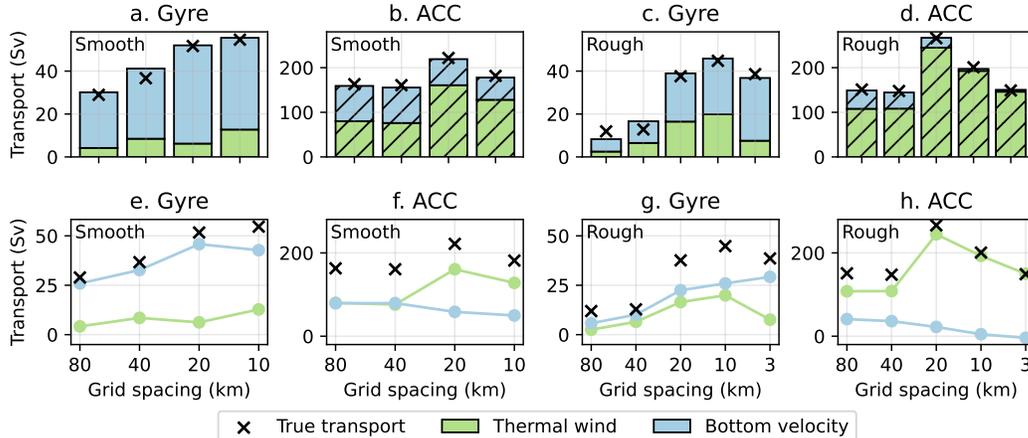


Figure 9. A decomposition of the gyre and ACC transports over smooth and rough bathymetries. The thermal wind component describes the geostrophic transport which emerges from horizontal density gradients. The bottom velocity component describes the depth-independent transport which is determined by the velocity at the sea floor ($u_b H$). The black crosses mark the total gyre and ACC transports shown in Figure 7.

4.4 Sensitivity of the thermal wind component to resolution

In the previous section we note that the depth-varying component of the flow can be closely described by the thermal wind relation and ultimately related to horizontal density gradients. To understand why the thermal wind component of the gyre and the ACC is particularly strong at eddy-permitting resolutions, we study the isopycnal structure at various resolutions. Zonal averages of the density are calculated over five meridional sections: Channel ($-2000 < x < -1000$ km), West Ridge ($500 < x < 1000$ km), East Ridge ($2000 < x < 2500$ km), East of the ridge ($3500 < x < 4000$ km), and the full zonal average. The locations of these sections are shown in Figure 6e.

Figure 11 compares the isopycnal structure between an eddy-permitting (10 km) and an eddy-parameterized (80 km) simulation with rough bathymetry. The isopycnal structure for simulations with a smooth bathymetry are qualitatively similar. In all meridional sections, the isopycnals are more tilted in the eddy-permitting simulation. In particular, meridional density gradients over the submarine ridge are large at eddy-permitting resolutions and the stratification of the gyre basin ($y < 0$) is reduced. At eddy-permitting and eddy-rich resolutions, the submarine ridge plays a large role in setting the nearby stratification; this is in agreement with observations (such as Figure 1b) and the findings of Wilson et al. (2022). From Figure 11 we can conclude that the thermal wind component of the gyre and ACC transports is larger in eddy-permitting models as density gradients are more extreme in the upper 2000 m of the model. A thick and weakly stratified layer also emerges in the eddy-permitting simulations, which is approximately below the 1028.5 kg m^{-3} contour. In this layer, horizontal density gradients are small and the thermal wind relation suggests that the zonal flow is not expected to vary significantly with depth.

Figure 12 compares the isopycnal structure between an eddy-rich (3 km) and the same eddy-permitting (10 km) simulation with rough bathymetry. The isopycnals of the eddy-rich and eddy-permitting simulations share similar features, however meridional density gradients are smaller in the eddy-rich case. This is particularly noticeable above the submarine ridge and ultimately reduces the thermal wind transport of the gyre and

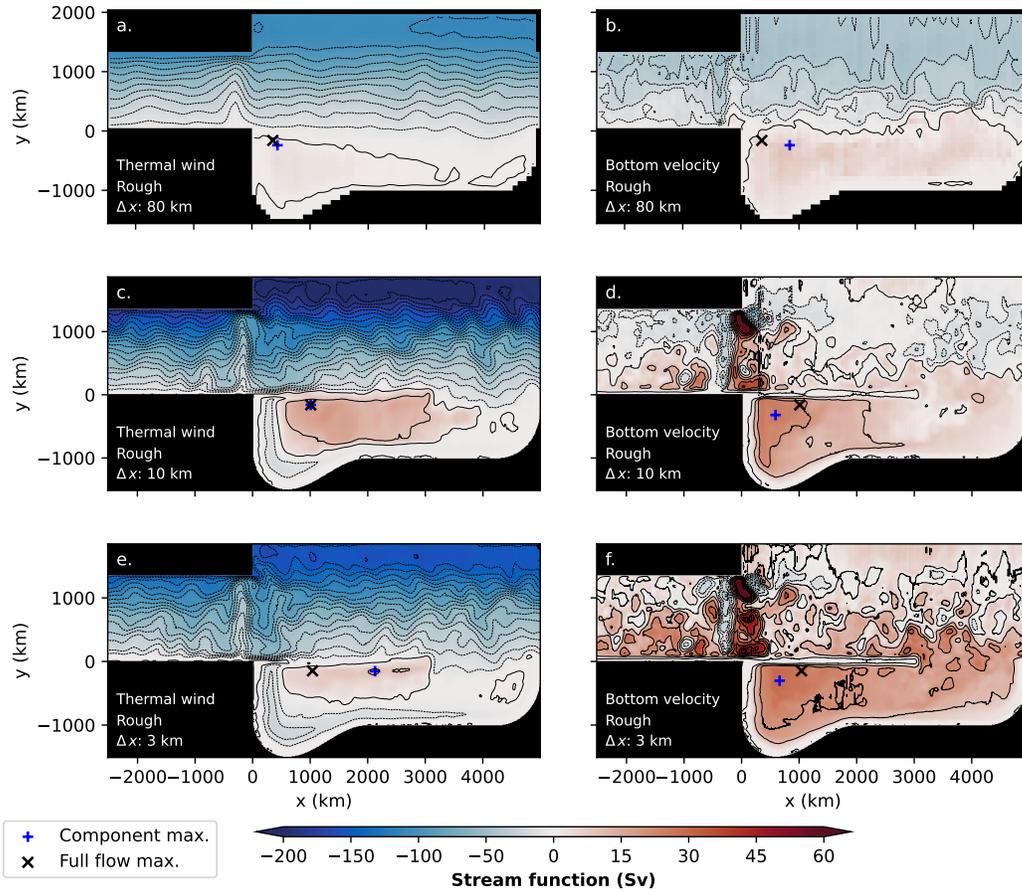


Figure 10. Stream functions from the same decomposition shown in Figure 9 at various resolutions. The black contours have an interval of 10 Sv. Results are from configurations with a rough bathymetry. The combined thermal wind and bottom velocity stream function is approximately equal to the total stream function, shown in Figure 7. Blue crosses mark the location of the component stream function maximum in the gyre basin. Black crosses mark the location where the total gyre transport is calculated.

ACC in the upper 2000 m of the model. The reduced outcropping of isopycnals in the eddy-rich model also increases the stratification of the gyre basin and reduces the thickness of the weakly stratified layer (approximately below 1028.5 kg m^{-3} contour). In simulations with explicit eddies, the densest isopycnals outcrop approximately 200 km south of the submarine ridge. In this area, the wind stress curl is at its peak negative value and there is little buoyancy forcing (see Figures 2c and d). It is therefore possible that there is an Ekman control on the maximum surface density in the gyre basin.

Such large horizontal density gradients are uncommon in eddy-parameterized models. The eddy-parameterized models (80 and 40 km) include the strongest tracer diffusion terms and eddy schemes like the GM parameterization simulate the effect of unresolved eddies by flattening isopycnals. In the presented simulations (and many climate models), the diffusion parameters and the parameterized eddies are insensitive to topographic features and consequently the isopycnals are relatively flat over the submarine ridge and the gyre basin. In reality, eddy diffusivity can be significantly reduced over large topographic features (Isachsen, 2011). This effect can be seen in our eddy-rich simulation as well; in Figure 6c, the eddy kinetic energy reduces by an order of magnitude above the submarine ridge and the continental shelf.

The eddy-permitting simulations are significantly less diffusive and no eddy-parameterization is used. The partially resolved eddy field is insufficient to flatten the isopycnals to the same extent as the eddy-parameterized models and therefore more extreme density gradients emerge. In the eddy-rich simulation (3 km), diffusive terms are small, but the near-resolved mesoscale eddy field is able to flatten the density surfaces more effectively than the partially-resolved eddy field. In Section 4.5, we test if various applications of GM in eddy-permitting models can change our results.

4.5 Using GM at eddy-permitting resolutions

In the results shown so far, all eddy-permitting simulations do not use the GM parameterization. In this subsection we investigate how strong to weak applications of GM affect our gyre and ACC transports at eddy-permitting resolutions. In simulations with 10 and 20 km horizontal grid spacings and rough bathymetry, the GM diffusivity parameter was varied from $35 \text{ m}^2\text{s}^{-1}$ up to $2000 \text{ m}^2\text{s}^{-1}$ (the value used in lower resolution simulations). The results of this sensitivity experiment are shown in Figure 13.

At both eddy-permitting resolutions (Figure 13a and 13d), the gyre transport decreases as the strength of the GM parameterization increases. Up until the strongest application of GM, the thermal wind component of the gyre transport weakens, suggesting the density surfaces are flattening in response to the increased GM diffusivity. When the GM diffusivity equals $2000 \text{ m}^2\text{s}^{-1}$, the residual of the thermal wind decomposition is significant, making it difficult to assess the relative importance of the thermal wind transport. The general weakening of the gyre transport is accompanied by a reduction in the overall eddy kinetic energy of the model (Figure 13c).

Figure 13f contextualises these eddy-permitting results. We find that weak applications of GM ($75 \text{ m}^2\text{s}^{-1}$ or less) in eddy-permitting models can improve their agreement with eddy-rich models at the expense of some explicit eddy kinetic energy. Stronger applications of GM ($600 \text{ m}^2\text{s}^{-1}$ or more) in eddy-permitting models can improve their agreement with the eddy-parameterized models but reduces the explicit mean eddy kinetic energy by an order of magnitude. In Figure 13f, the gyre transport still slowly increases with resolution when a GM diffusivity of $2000 \text{ m}^2\text{s}^{-1}$ is used. In the absence of explicit eddies, this slow increase is likely a result of increased resolution of the mean flow, increased resolution of the bathymetry, or the linear variation of the momentum and tracer diffusivity with horizontal grid spacing.

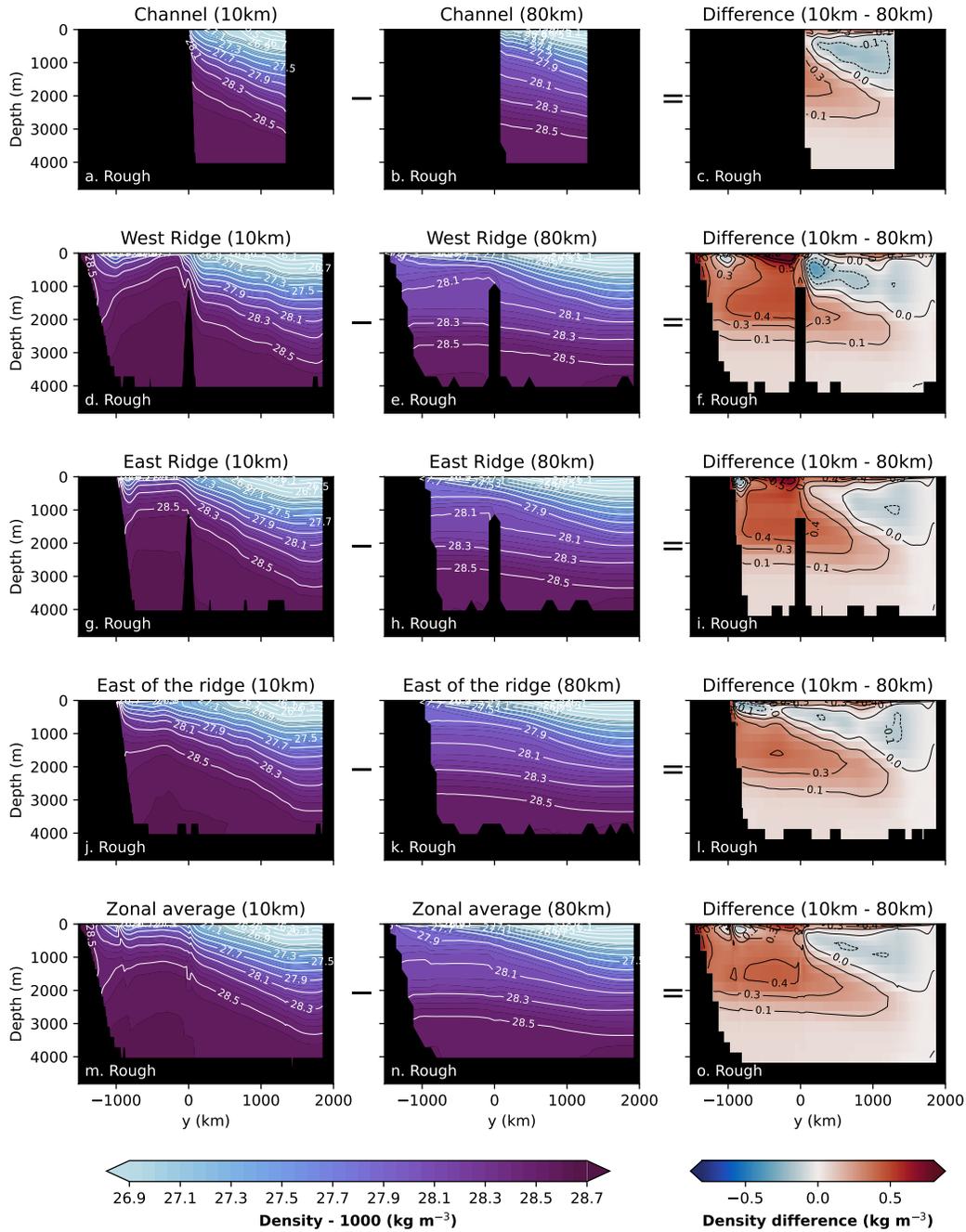


Figure 11. Meridional density sections of an eddy-permitting (10 km, left column) and an eddy parameterized (80 km, middle column) simulation. The difference between the density sections are shown in the right column. All presented sections feature a rough bathymetry.

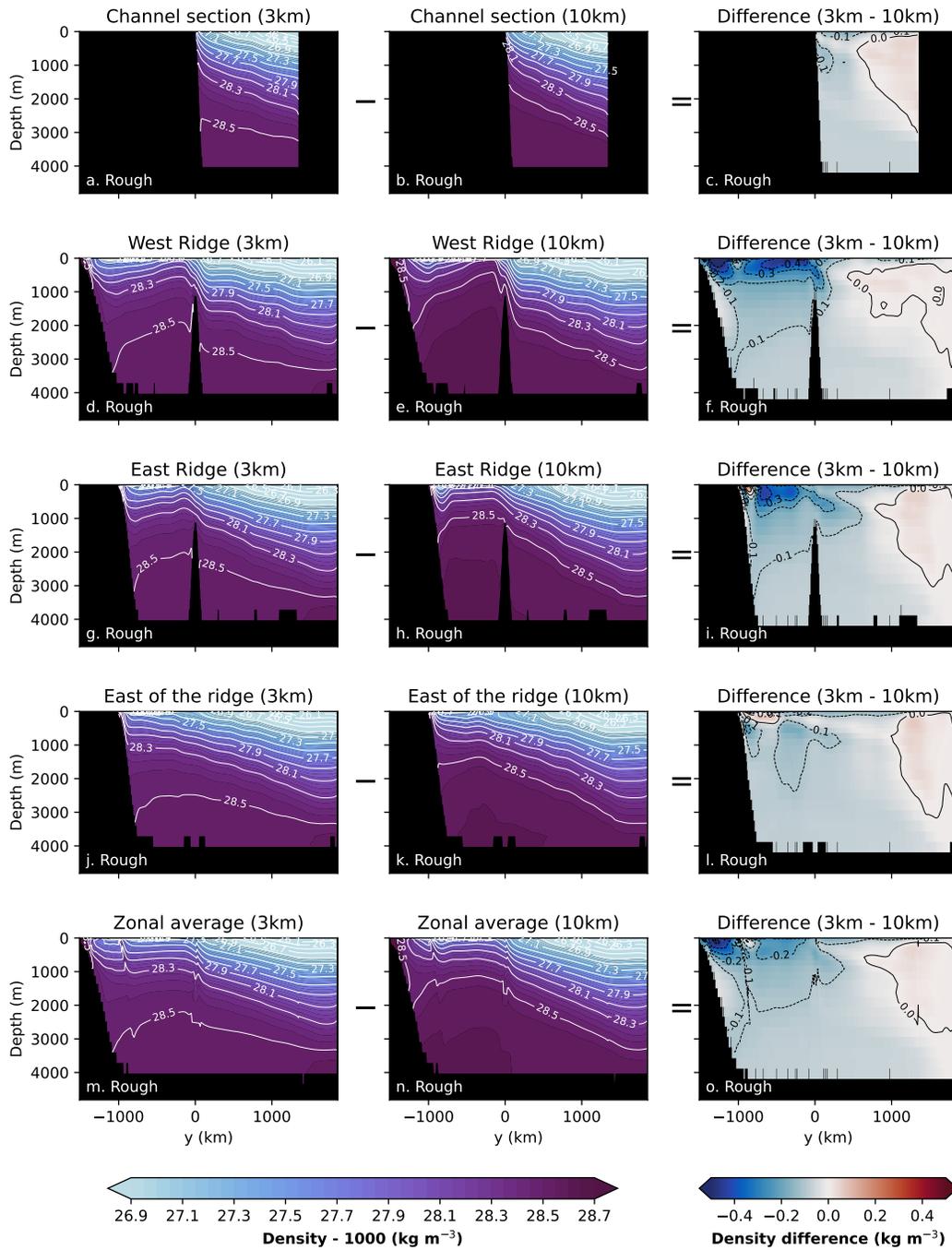


Figure 12. Meridional density sections of an eddy-rich (3 km, left column) and an eddy-permitting (10 km, middle column) simulation. The difference between the density sections are shown in the right column. All presented sections feature a rough bathymetry.

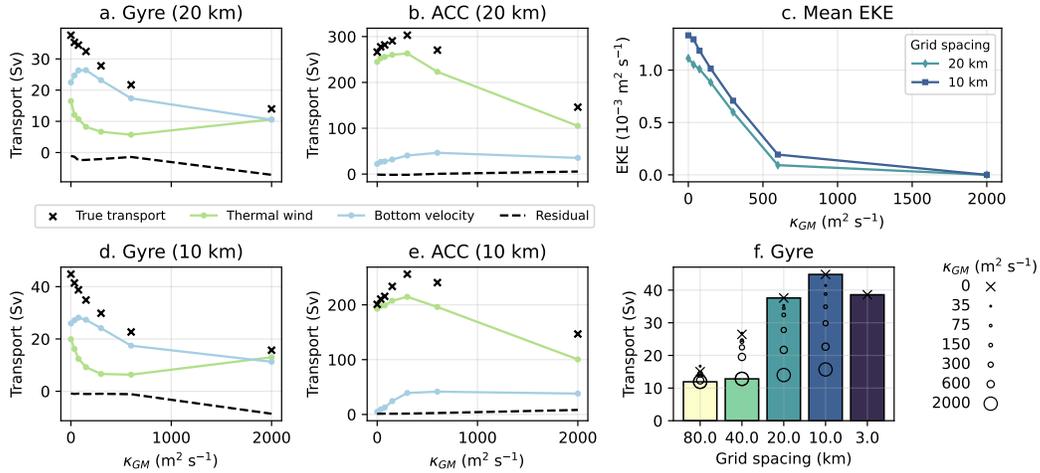


Figure 13. The effect of using GM in eddy-permitting models. (a) and (d) show the transport of the Weddell Gyre with applications of GM at various strengths alongside the thermal wind decomposition. (b) and (e) show the transport of the ACC with applications of GM at various strengths alongside the thermal wind decomposition. (c) is the mean eddy kinetic energy when the GM diffusivity parameter is varied. (f) compares the various gyre transports when GM is applied to the transports shown in Figure 7c. The strength of GM is similarly varied in the eddy-parameterized models and the effect on the gyre transport is shown in (f).

502 For completeness, we also shown in Figure 13f what happens to the gyre transport
 503 when the GM parameter is similarly varied in eddy-parameterized models. In the 80 km
 504 case, the gyre transport was insensitive to variations in the GM parameter and has a max-
 505 imum value of 16.7 Sv when the GM parameter is $35 m^2 s^{-1}$. In the 40 km case, the gyre
 506 transport is more sensitive to the variation in the GM parameter and has a maximum
 507 value of 26.8 Sv when the GM paramater is $35 m^2 s^{-1}$. Significantly reducing the strength
 508 of GM in eddy-parameterized models (from 2000 to $75 m^2 s^{-1}$ or less) can increase the
 509 gyre transport but does not lead to agreement with the eddy-rich simulations.

510 Figure 13f shows that in our experiment, no selection of GM diffusivities can pro-
 511 duce a consistent gyre transport between all horizontal resolutions. Using different ap-
 512 plications of GM (e.g. spatially varying diffusivities) and altenative parameterizations
 513 may yield different results. Exploring these approaches with the model would be a valu-
 514 able contribution to the ocean modelling community but is beyond the scope of the present
 515 study.

516 5 Discussion

517 In the previous section, we related the thermal wind component of the gyre and
 518 ACC transports to the density structure. In contrast, it is not immediately clear why
 519 the transport from the bottom flow can vary significantly with resolution. To explore
 520 this sensitivity of the bottom flow to resolution, we discuss the potential contributions
 521 of non-linear eddy-mean flow interactions. We will then discuss other physical processes
 522 that are neglected in our idealized model.

5.1 Role of eddies

In this article, we observe that the density surfaces of the idealized Weddell Gyre change significantly when explicit eddies are introduced to ocean simulations and that the density structure has a significant influence on the horizontal circulation. In the presence of rough bathymetry, the thermal wind component of the gyre transport intensifies in simulations with explicit eddies because the meridional density gradients are steeper. In Figure 6, we see large eddy kinetic energies in areas where the isopycnal tilt is the most extreme. This suggests that the increase in the mean available potential energy that comes with steeper isopycnals can fuel more energetic mesoscale eddies.

A component of the mesoscale eddy field extends to the sea floor, as shown by the vertical bands of eddy kinetic energy in Figure 6. It therefore seems likely that explicit mesoscale eddies are influencing the bottom flow in select areas of the model (e.g. between 1000 and 500 km south of the submarine ridge in Figure 6b). In the presence of variable bottom topography, mesoscale eddies can drive a mean circulation along topographic contours (Bretherton & Haidvogel, 1976). This can be interpreted either as an ‘entropic force’ (Holloway, 1987, 1992) or, alternatively, as energetically constrained mixing of potential vorticity over the sloping topography (Bretherton & Haidvogel, 1976; Adcock & Marshall, 2000). These eddy-driven flows are not captured by eddy parameterizations employed in climate models based on GM. However, it is possible that these entropic forces are captured in our simulations with explicit eddies, consistent with the strengthening of the bottom flow at higher resolutions. This deserves further investigation, but is beyond the scope of the present study.

As eddy-permitting models become more feasible for climate projections, there is an increasing interest in developing eddy parameterizations for simulations where the largest eddies are at least marginally resolved. The development and testing of eddy parameterizations is a busy area of research; Hewitt et al. (2020) review the various approaches that could be deployed in eddy-parameterized and eddy-permitting ocean models. Of particular relevance to the Weddell Gyre is a recent study by Wei et al. (2022) which finds encouraging results for parameterized mesoscale eddy buoyancy fluxes over large scale bathymetry when topographic suppression effects are incorporated.

5.2 Missing physics in the idealized configuration

Before concluding on the results presented in this article, it is important to summarize the limitations of the idealized model. Firstly, the winds in these configurations are zonal and do not follow the continental shelf. This may be the reason why the density driven slope current in this model does not reach the Drake Passage; additionally, a complete slope current may require deep passages in the zonal submarine ridge. In Figure 12, the isopycnals tilt upwards when above the continental shelf. This suggests that dense water is being exported everywhere along the shelf. In reality, dense water export is more localized and isopycnals would tilt down above the continental shelf in East Antarctica (Thompson et al., 2018). Non-linearities in the equation of state may also be important for AABW formation (Gill & Niller, 1973), so using a linear equation of state may lead to inaccurate densities. Presently, it is unclear if a more accurate slope current would modify the gyre transport significantly.

We imitate the time-averaged effect of ice with an effective freshwater flux shown in Figure 2d but no attempt is made to couple the effect of sea ice to the oceanic or atmospheric state. In addition, the effect of internal stresses in the ice can modify the surface stress experienced by the ocean. By neglecting internal stresses, we are assuming that all ice is in ‘free drift’ which may not be valid near the continental shelf according to satellite observations (Kimura, 2004; Kwok et al., 2017). In reality, the Weddell Gyre and ACC are exposed to an extreme seasonal cycle. The seasonal variability of the wind stress and buoyancy forcing can be larger than the time-averaged forcing. In this work

574 we are assuming that time-averaged forcing will accurately produce a time-averaged Wed-
 575 dell Gyre and ACC, but this may not be true because of non-linear processes. This is
 576 certainly not true for the subpolar gyres in the northern hemisphere, where winter con-
 577 ditions play a disproportionately large role in setting the properties of the deep ocean
 578 thermocline as waters subducted at any time outside of late-winter are re-entrained by
 579 the dynamic mixed layer (Stommel, 1979). A similar mechanism also operates on an inter-
 580 annual time scale in the northern hemisphere (MacGilchrist et al., 2021). It is unclear
 581 if a similar selection process (‘Stommel’s Demon’) is present in the Southern Ocean and
 582 needs further investigation. All experiments used in this study are in a statistically steady
 583 state (see Figure 4), unlike the real ocean which is exposed to an extreme seasonal cy-
 584 cle, a changing global ocean, and a changing climate.

585 The large-scale topographic features in the model (shown in Figure 2a) are qual-
 586 itatively similar to the Weddell basin but there are some important differences. Firstly,
 587 the submarine ridge and the northern boundary of the domain are zonal. The meridional
 588 components of the idealized Weddell Gyre and ACC are too constrained by bottom top-
 589 ography when compared to the real ocean. In reality, the ACC is deflected northwards
 590 immediately upon exiting the Drake Passage (Mazloff et al., 2010) which is a behaviour
 591 this idealized model cannot recreate as any northward deflection of the ACC is limited
 592 by the sponge region. Finally, a unique feature of the Weddell Gyre is its dynamic shape
 593 as no obvious topographic feature constrains the gyre’s eastern boundary (Vernet et al.,
 594 2019). In our idealized simulations, the zonal extent of the Weddell Gyre is not able to
 595 extend beyond the width of the basin (5000 km) without taking a northward departure
 596 into the ACC channel.

597 6 Conclusions

598 Using a minimal description of the Weddell Gyre and ACC, we have identified an
 599 extreme sensitivity of the circulation to horizontal grid spacing between eddy-parameterized
 600 and eddy-permitting resolutions. The Weddell Gyre in eddy-permitting simulations is
 601 significantly stronger than in eddy-parameterized cases and slightly stronger than an eddy-
 602 rich case (Figure 7). This is concerning as coupled climate models are beginning to tra-
 603 verse this highly sensitive ‘gray zone’, where large mesoscale eddies are only partially re-
 604 solved.

605 To investigate if the gyre transports are affected by the varying ACC strength, we
 606 performed a sensitivity experiment in Section 4.2. The channel topography was mod-
 607 ified to either increase or decrease the ACC transport at eddy-parameterized and eddy-
 608 permitting resolutions and the effect on the Weddell Gyre was negligible (Figure 8). The
 609 limited coupling of the gyre and ACC transports was useful for our study but the insen-
 610 sitivity itself is also scientifically interesting and should be investigated further.

611 To improve our understanding of the flow’s vertical structure, we used a thermal
 612 wind decomposition in Section 4.3 which works well with a small residual (Figure 9). In
 613 cases with a smooth bathymetry, the gyre strength is almost entirely determined by the
 614 depth-independent, bottom flow transport, $\mathbf{u}_b H$. When a rough bathymetry is used, the
 615 bottom gyre transport is comparable in size to the thermal wind transport, which varies
 616 with depth. Although the total transport sensitivity to resolution is similar with smooth
 617 and rough bathymetry, the vertical and horizontal structure of the flow clearly differs.
 618 This highlights how permitting small topographic interactions everywhere in an ideal-
 619 ized model can change large scale circulation features.

620 With a rough bathymetry, the thermal wind component of the gyre is particularly
 621 large over the submarine ridge (Figure 11 and 12). This is a behaviour which was found
 622 in the work of Wilson et al. (2022) and in hydrographic sections (such as Figure 1b). Wilson
 623 et al. (2022) found that the submarine ridge plays a fundamental role in setting the strat-

624 ification and circulation of the Weddell Gyre and the ACC. This result holds true in our
 625 experiments at eddy-permitting and eddy-rich resolutions. However, we would need to
 626 study simulations without the submarine ridge to fully assess its impact on the gyre at
 627 various resolutions.

628 The thermal wind component of the gyre is largest at eddy-permitting resolutions
 629 because the partially-resolved eddy field is less effective at flattening isopycnals than a
 630 fully-resolved eddy field or an eddy parameterization. In all cases, the bottom flow trans-
 631 port of the gyre increases significantly when explicit eddies are present. In Section 4.5,
 632 we showed that using a weak application of the GM parameterization can bring eddy-
 633 permitting gyre transports down to eddy-rich values at the expense of some explicit eddy
 634 kinetic energy (Figure 13). Only a strong application of GM can significantly reduce the
 635 gyre’s extreme sensitivity to horizontal resolution between eddy-parameterized and eddy-
 636 permitting scales and this would be at the expense of practically all explicit eddy kinetic
 637 energy. We also varied the strength of GM in the eddy-parameterized models and found
 638 that no selection of GM parameters can produce a consistent gyre transport across all
 639 horizontal resolutions.

640 In this study, the Weddell Gyre transport is largest and the isopycnals are the steep-
 641 est at eddy-permitting resolutions. For this reason, ocean modellers should approach this
 642 eddy-permitting ‘gray zone’ with care when simulating the Southern Ocean and consider
 643 employing parameterizations that are compatible with partially resolved mesoscale ed-
 644 dies.

645 **Appendix A Computational cost of the experiments**

646 In this section, we summarize the computational cost of running the idealized model
 647 at various resolutions. On ARCHER2 the 20 km model required 1.2 node hours per model
 648 year; the 10 km model required 6.1 node hours per model year; and the 3 km model re-
 649 quired 154.0 node hours per model year. On Monsoon2, the 80 km model required 0.4
 650 node hours per model year and the 40 km model required 0.1 node hours per model year.
 651 The 80 km model has 132,525 grid points; the 40 km model has 509,733 grid points; the
 652 20 km model has 1,998,477 grid points; the 10 km model has 7,889,934 grid points; and
 653 the 3 km model has 87,077,760 grid points.

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 661 UK collaborative data analysis facility. Data were collected and made publicly available
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666 **Open Research**

667 Input files for the idealized model integrations are archived on Zenodo (Styles et al., 2023b).
 668 The idealized configuration, designed for NEMO release 4.0.1, is available on Zenodo (Styles
 669 et al., 2023c). The software used to analyse the model outputs is also archived on Zen-
 670 odo (Styles et al., 2023a).

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Figure 1.

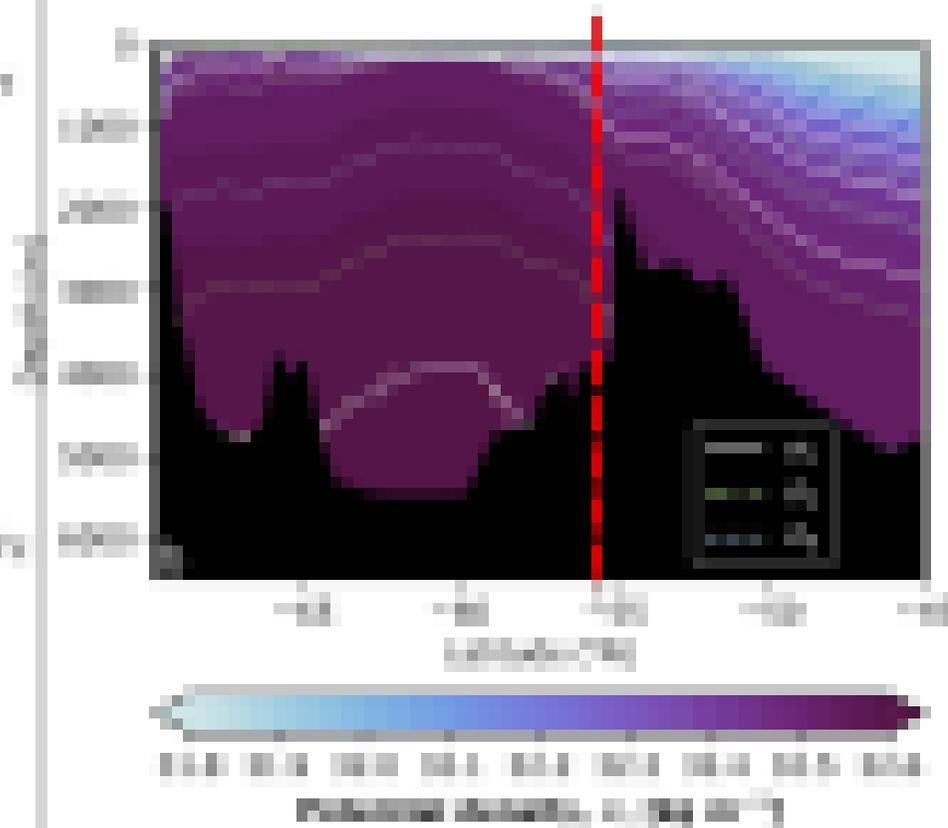
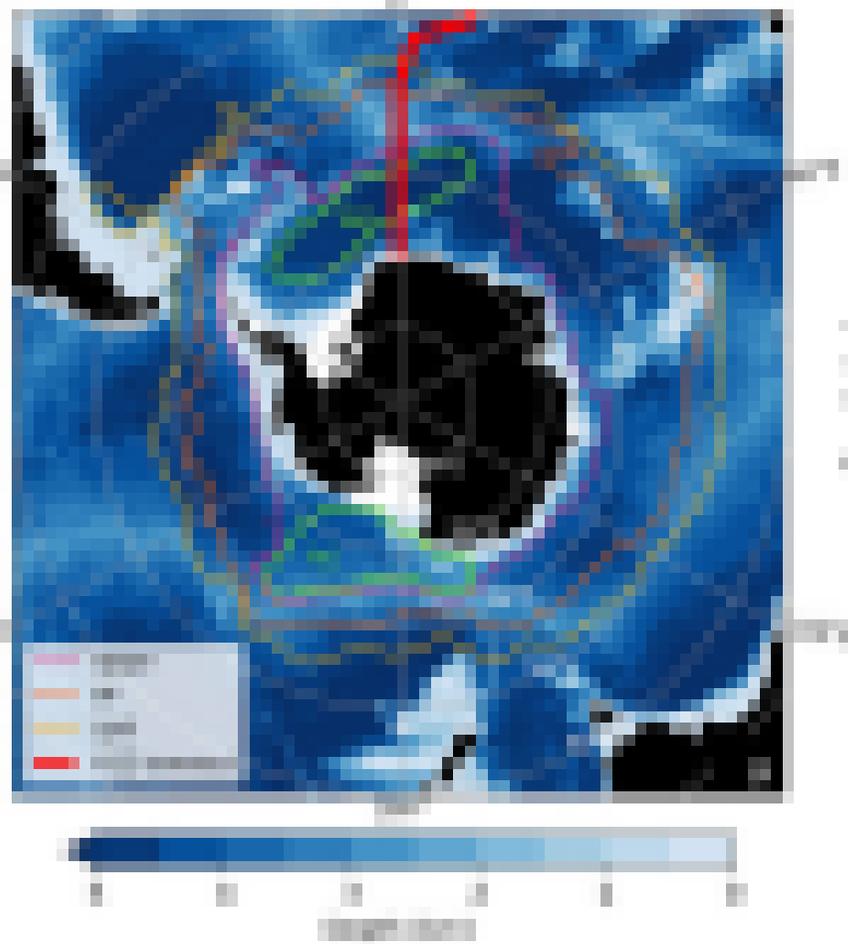
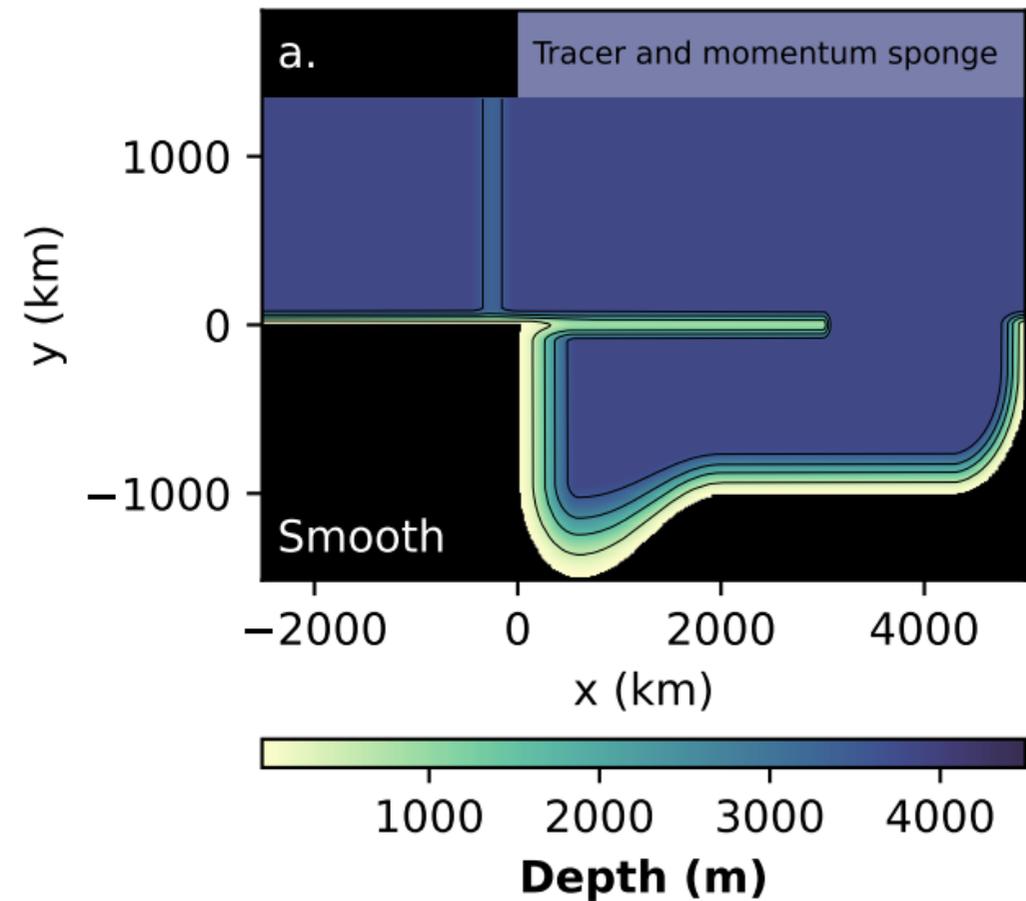
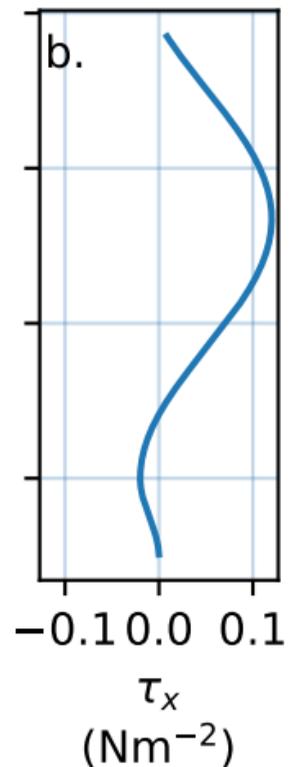


Figure 2.

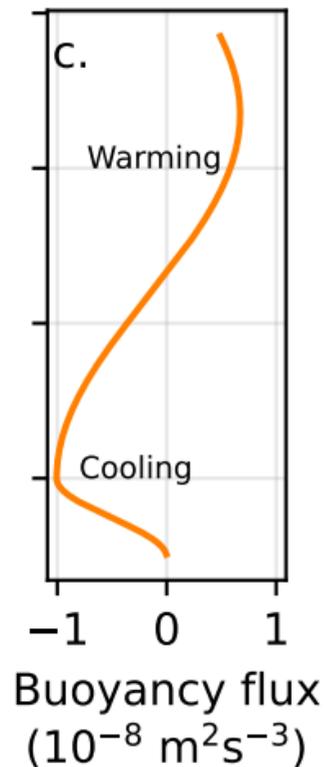
Bathymetry



Wind stress



Heat



Idealized ice

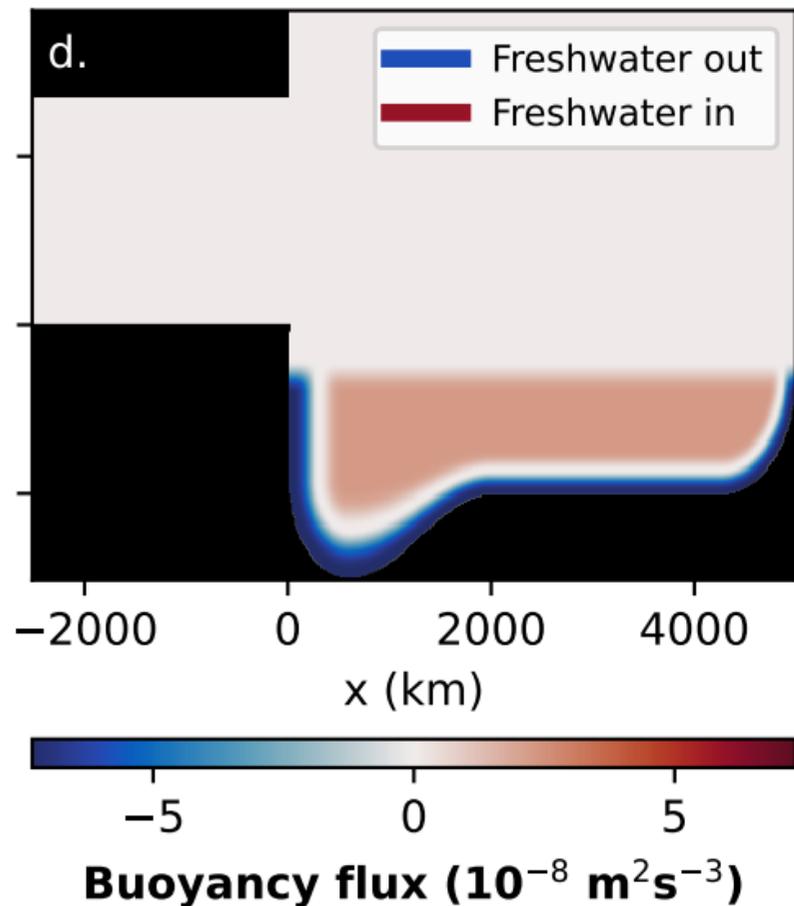


Figure 3.

Topographic noise

Bathymetry

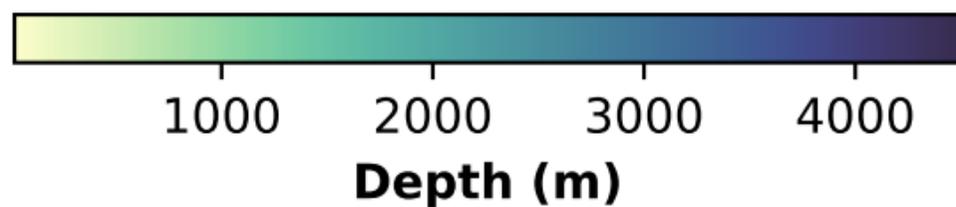
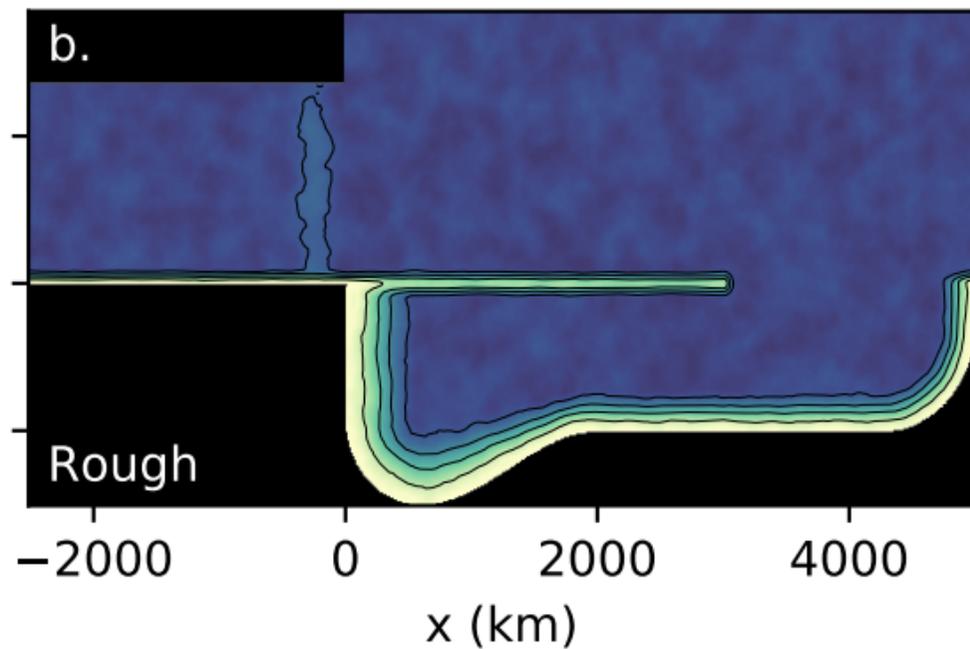
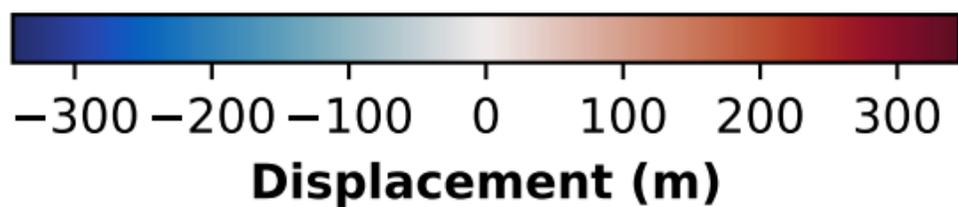
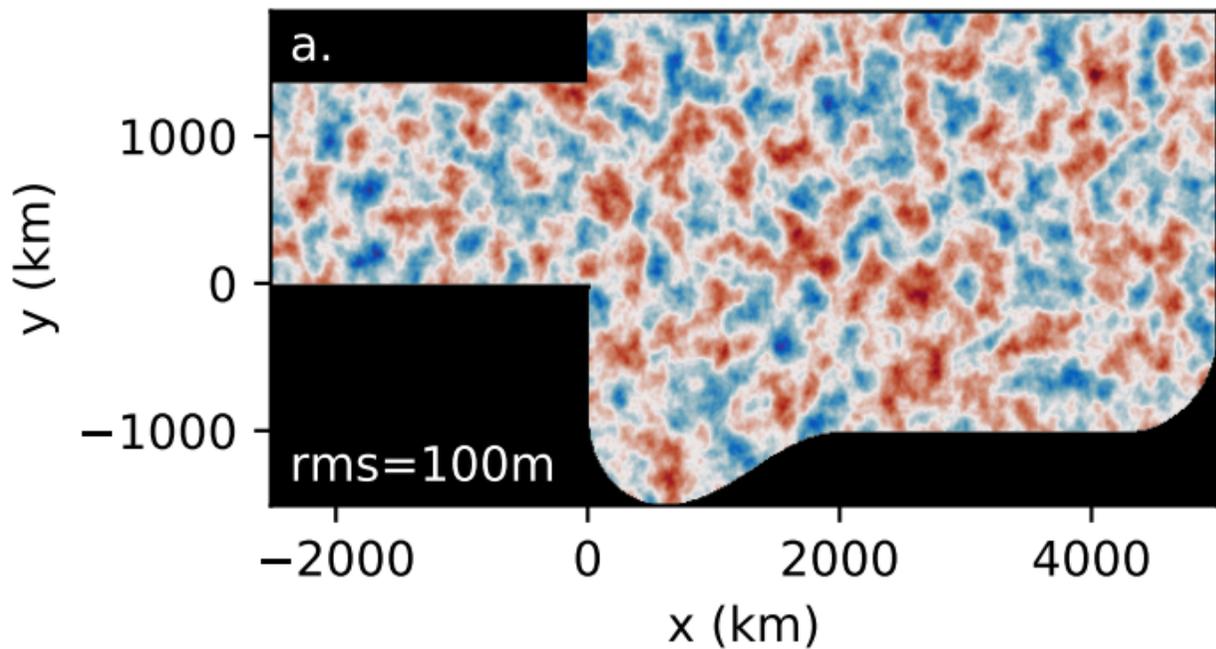


Figure 4.

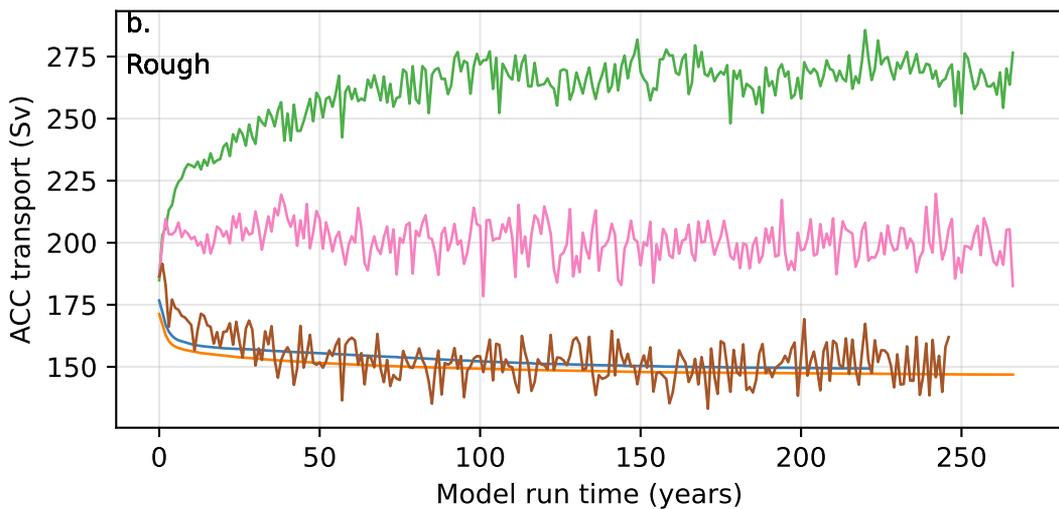
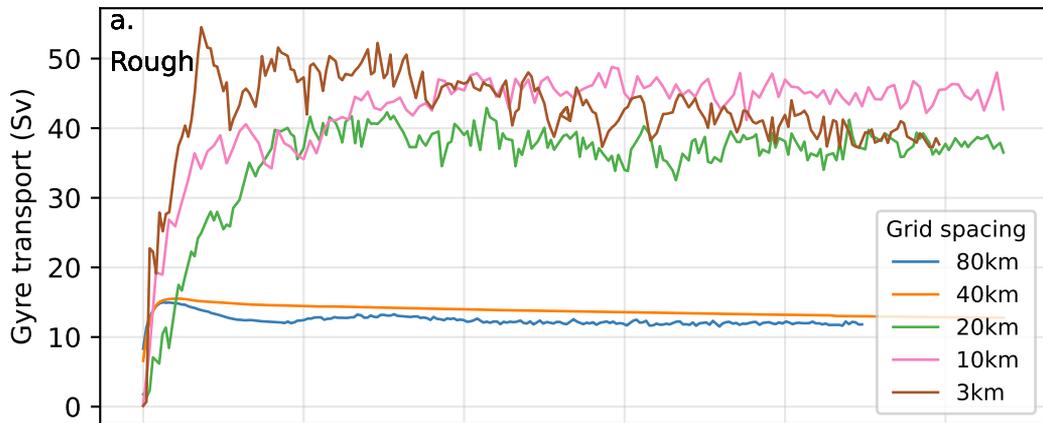


Figure 5.

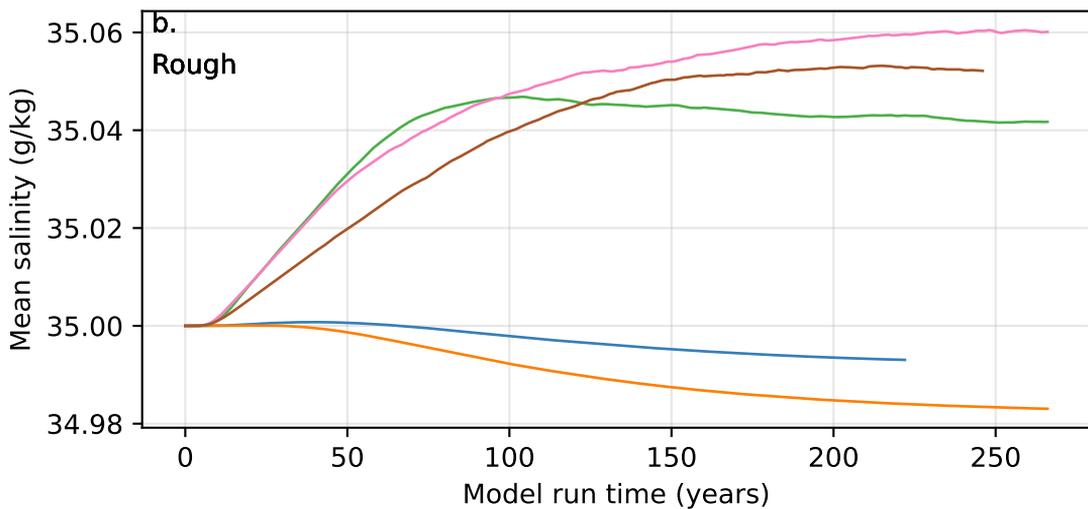
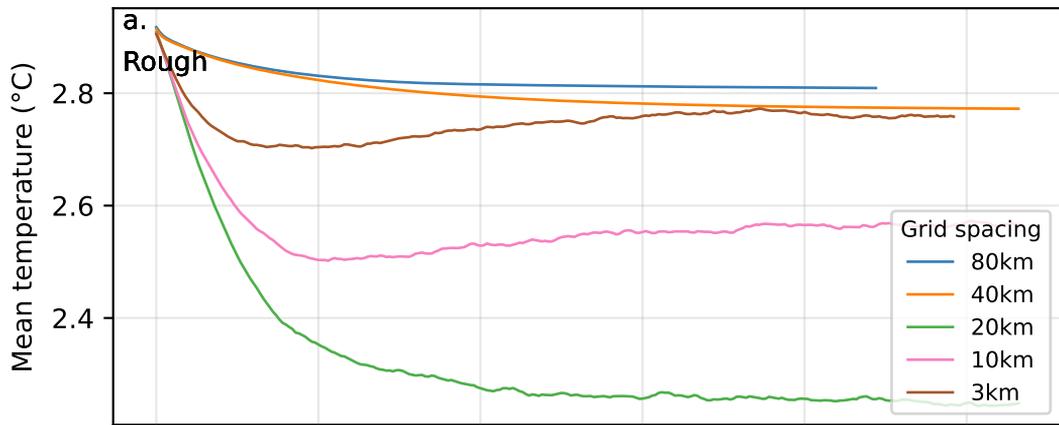
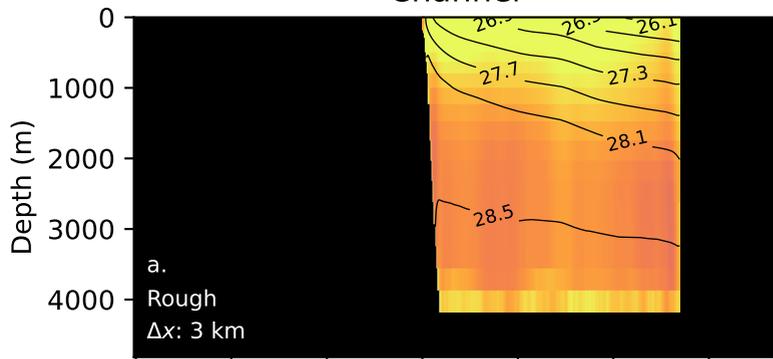
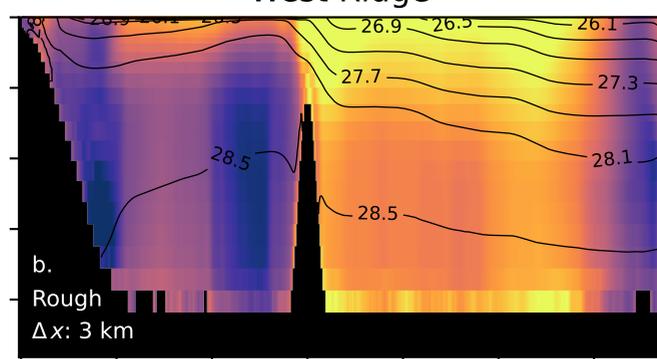


Figure 6.

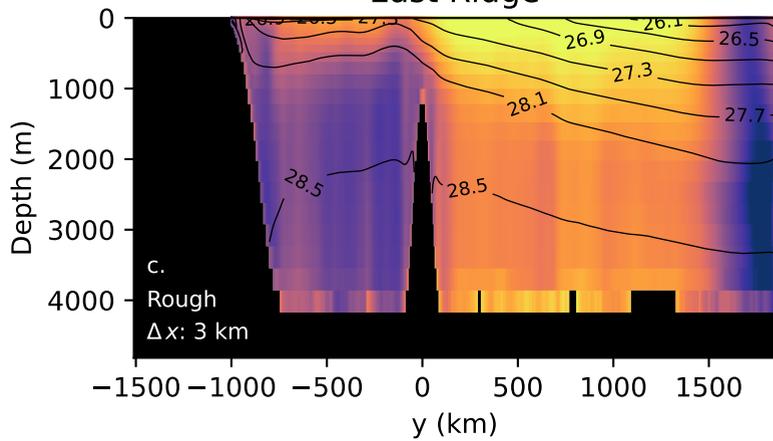
Channel



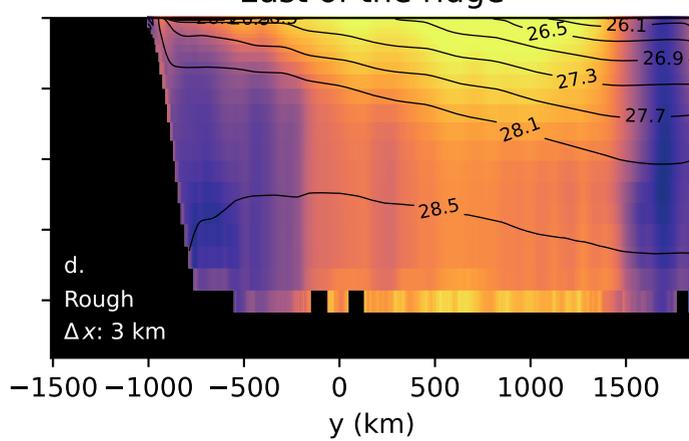
West Ridge



East Ridge



East of the ridge



Depth-averaged EKE

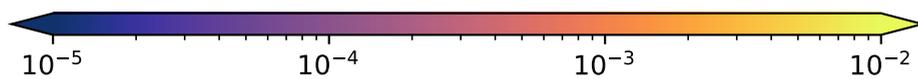
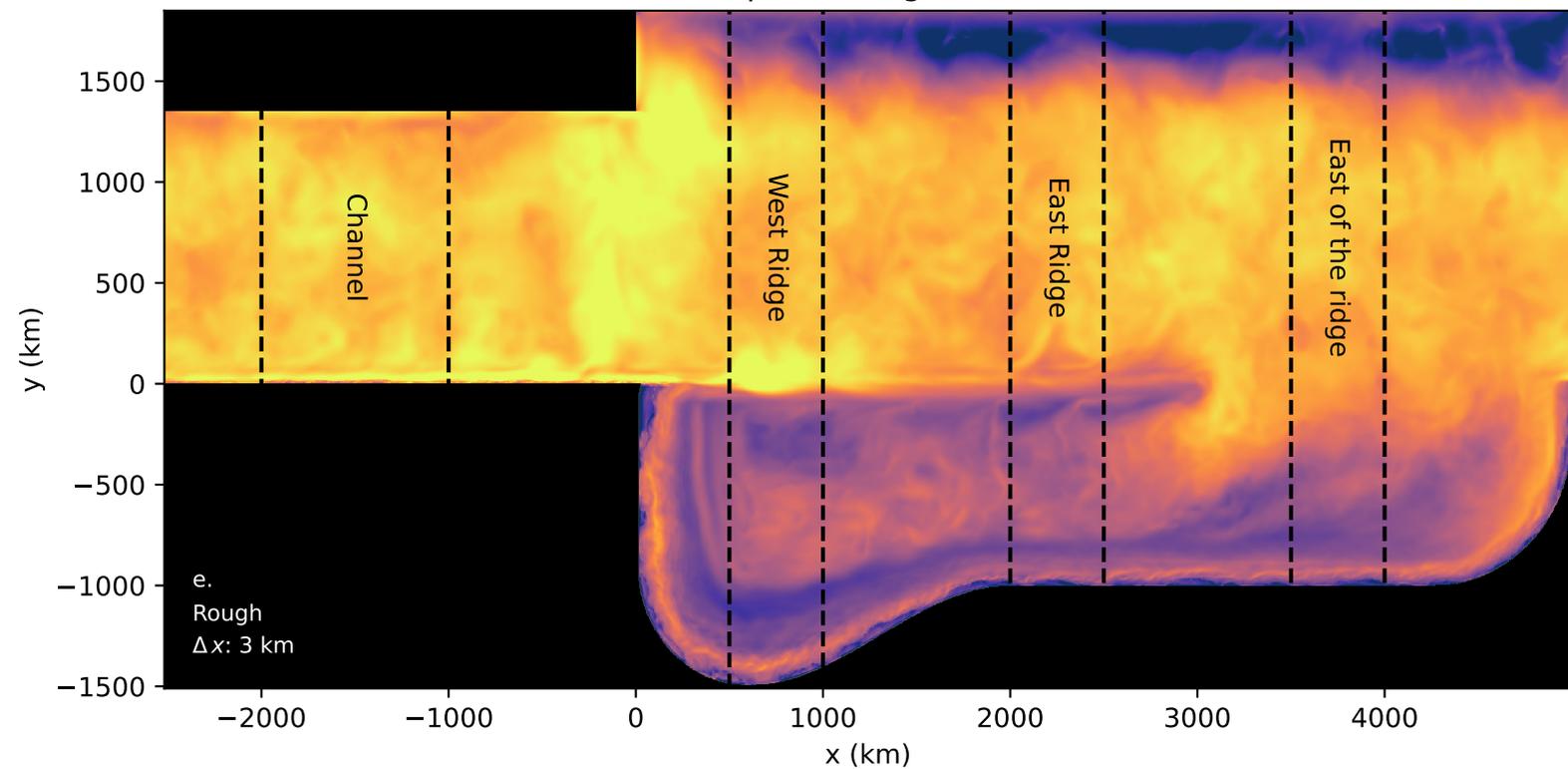
Eddy Kinetic Energy (m^2s^{-2})

Figure 7.

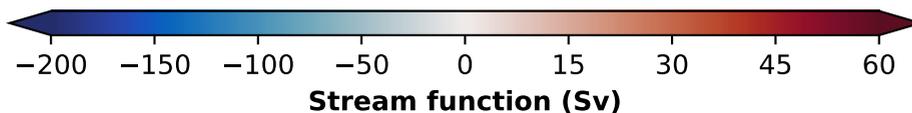
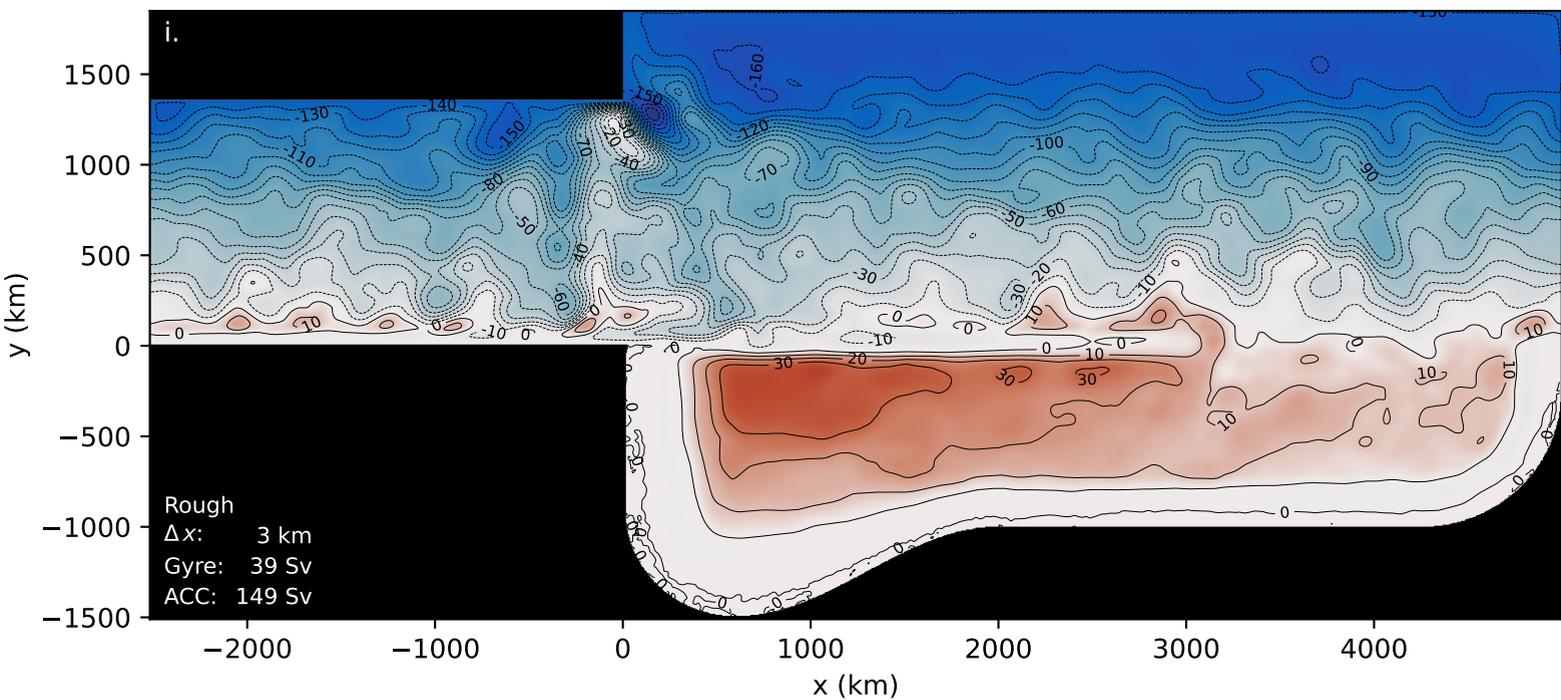
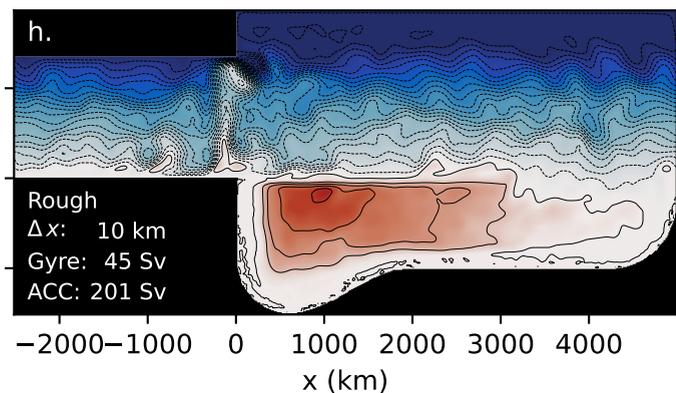
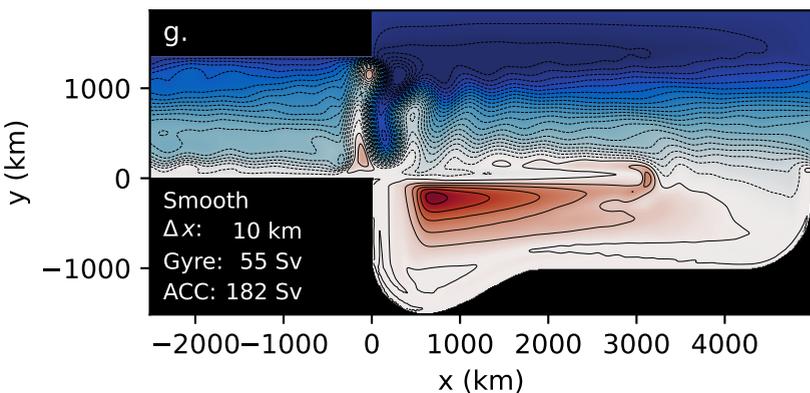
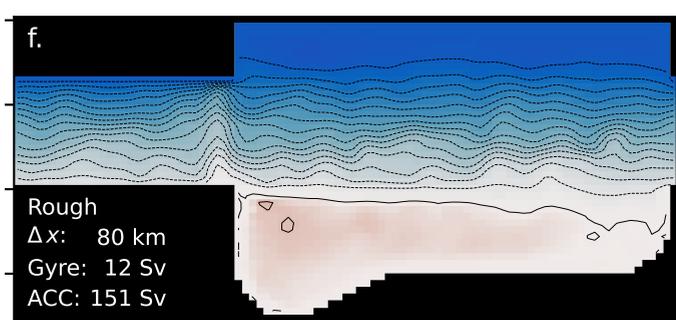
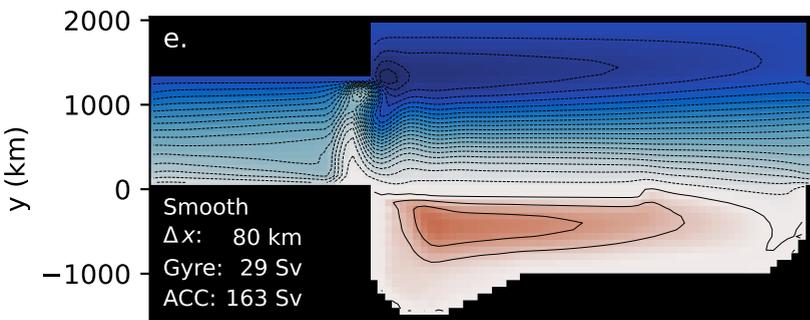
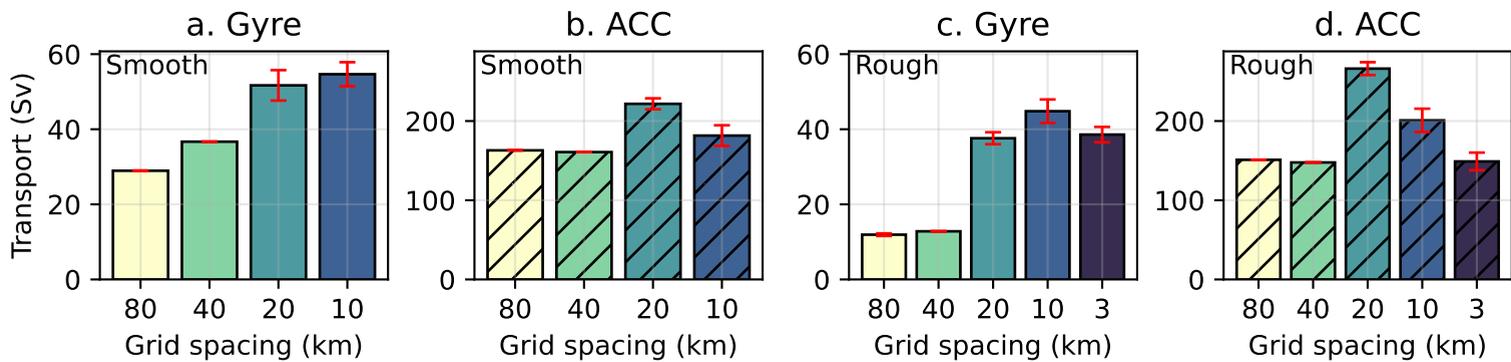


Figure 8.

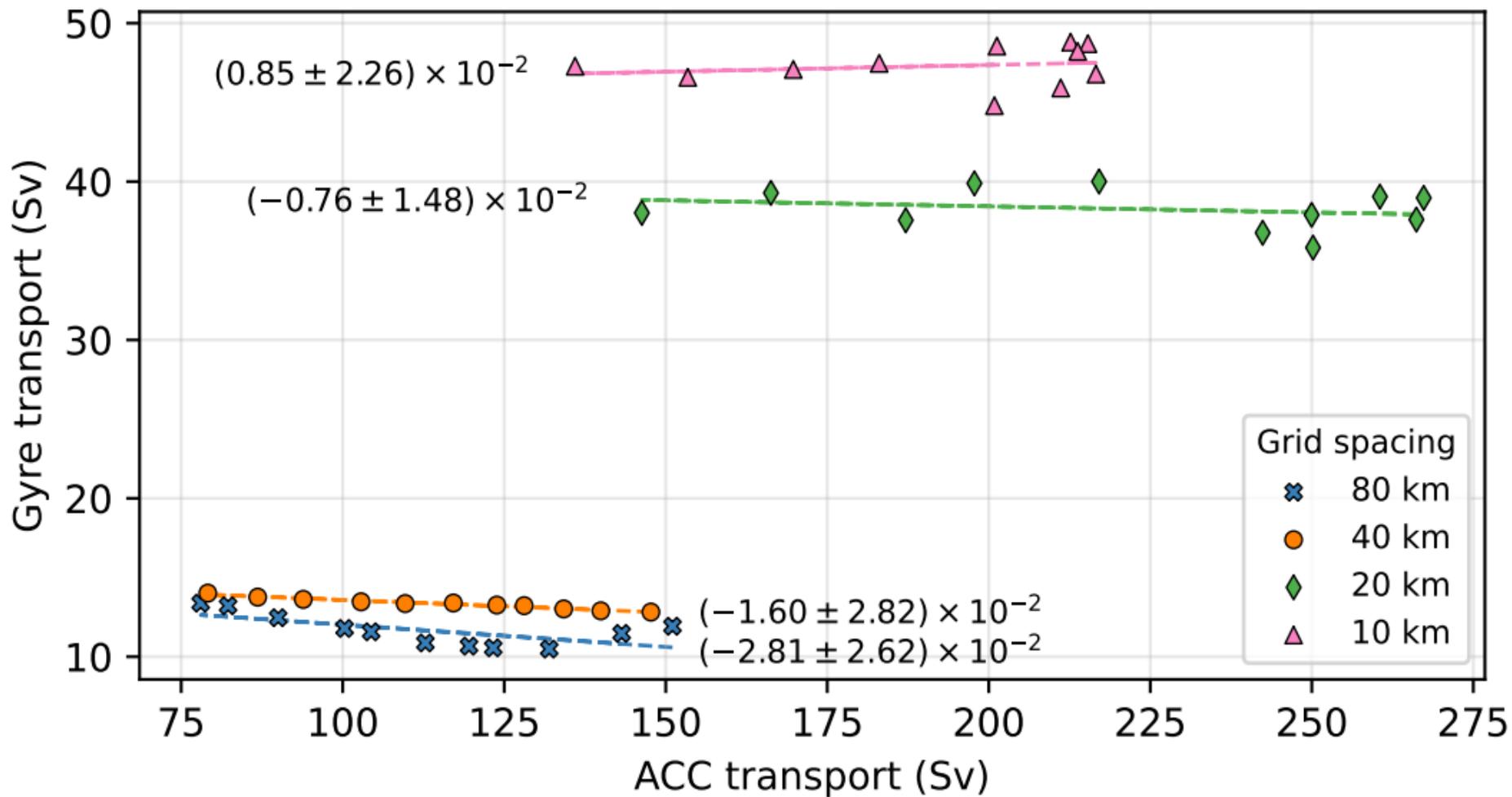
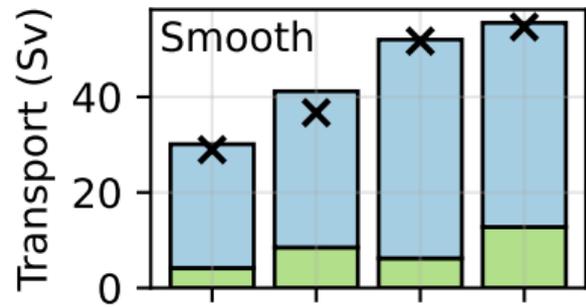
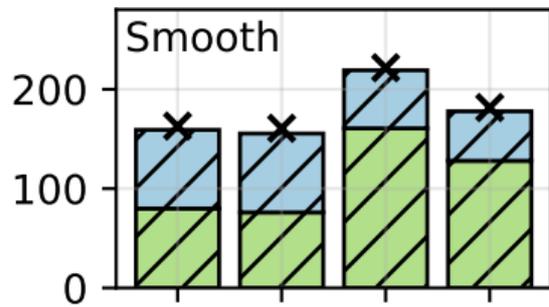


Figure 9.

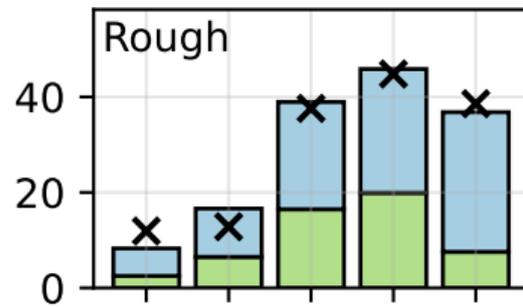
a. Gyre



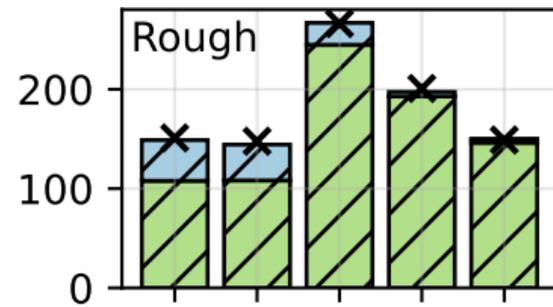
b. ACC



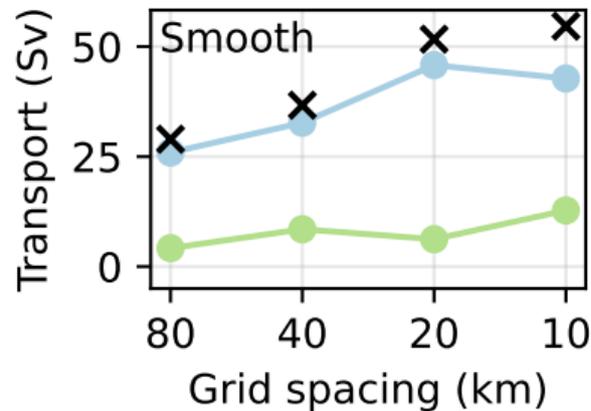
c. Gyre



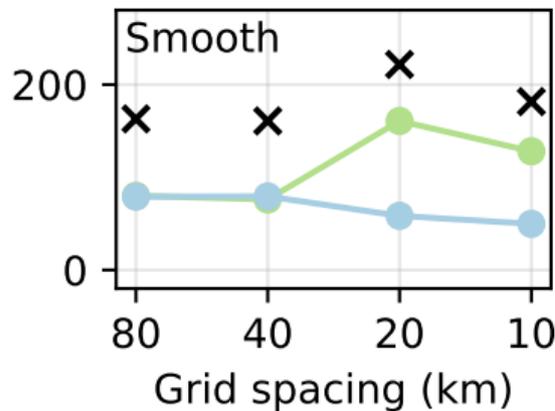
d. ACC



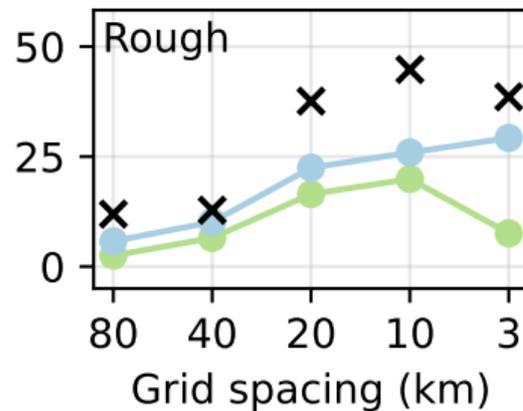
e. Gyre



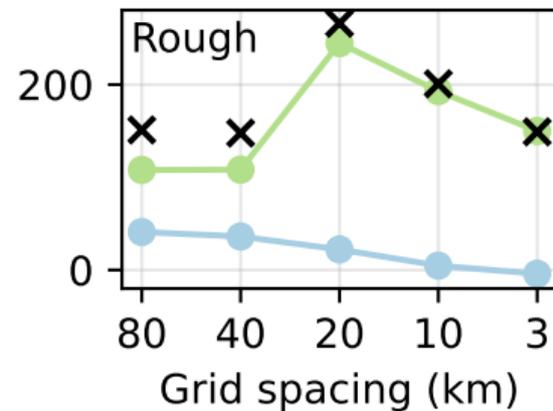
f. ACC



g. Gyre

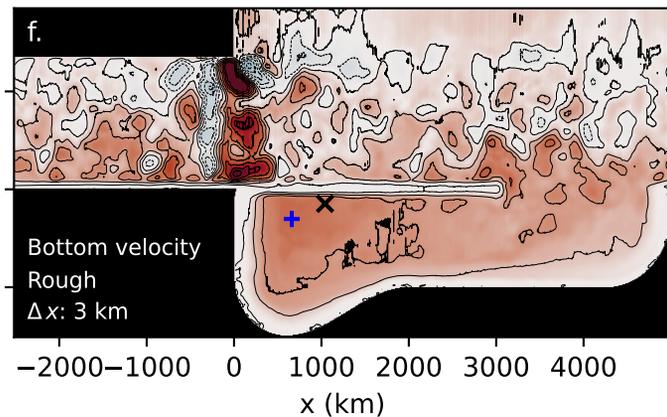
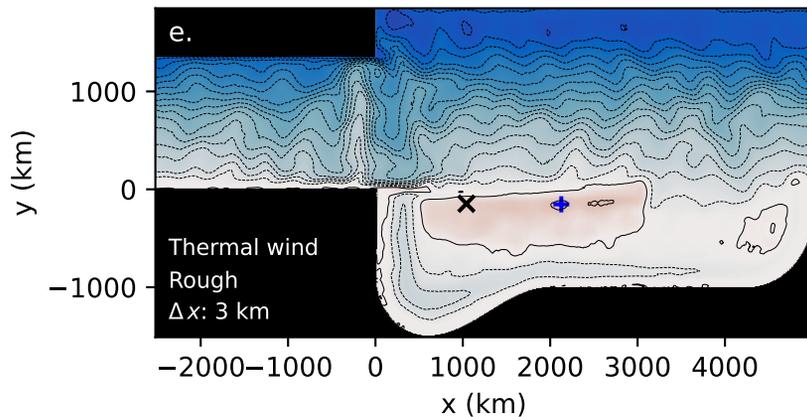
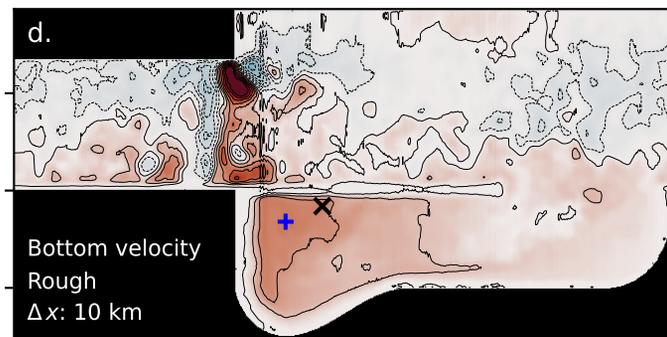
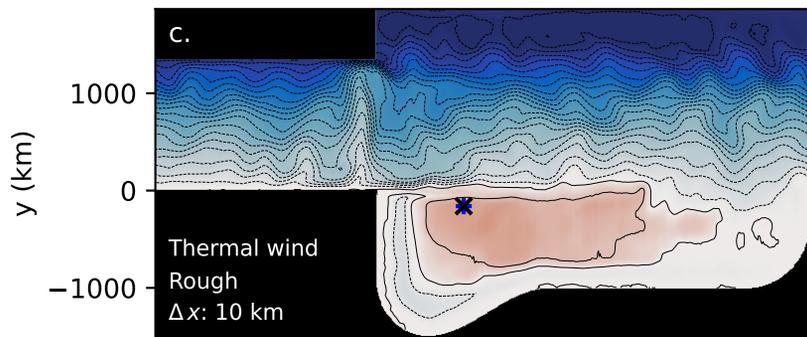
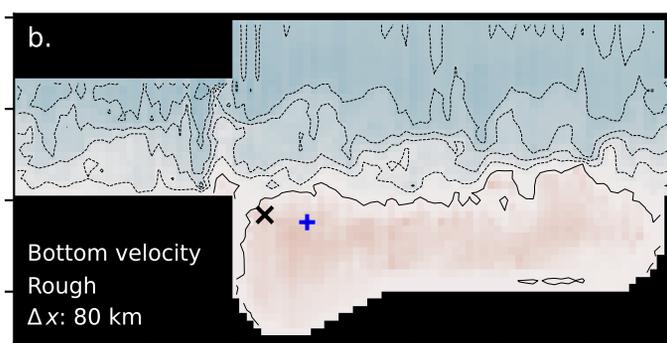
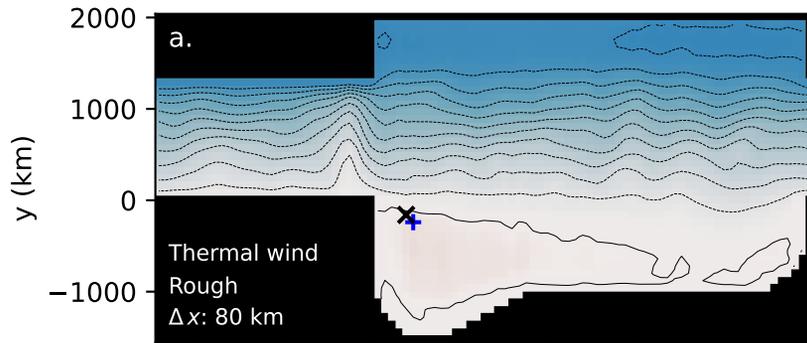


h. ACC



x True transport
 Thermal wind
 Bottom velocity

Figure 10.



+ Component max.
x Full flow max.

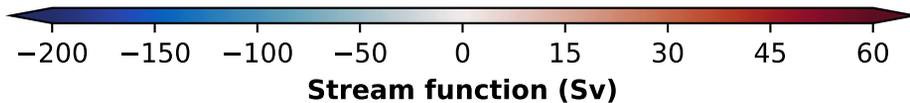
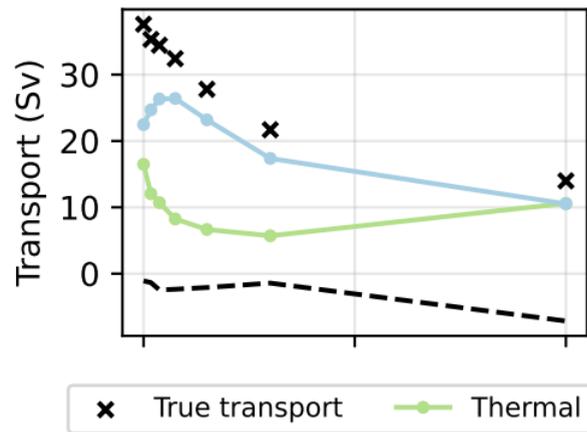


Figure 11.

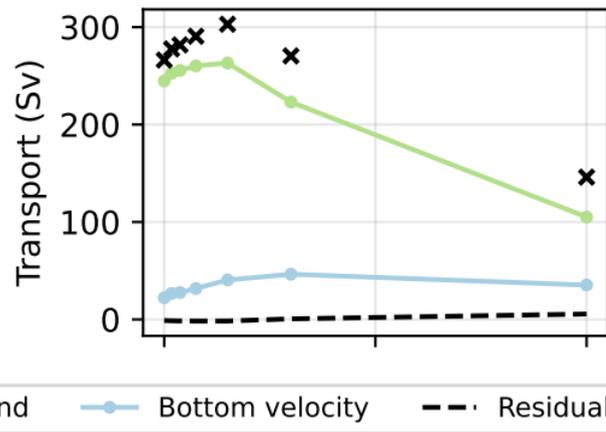
Figure 12.

Figure 13.

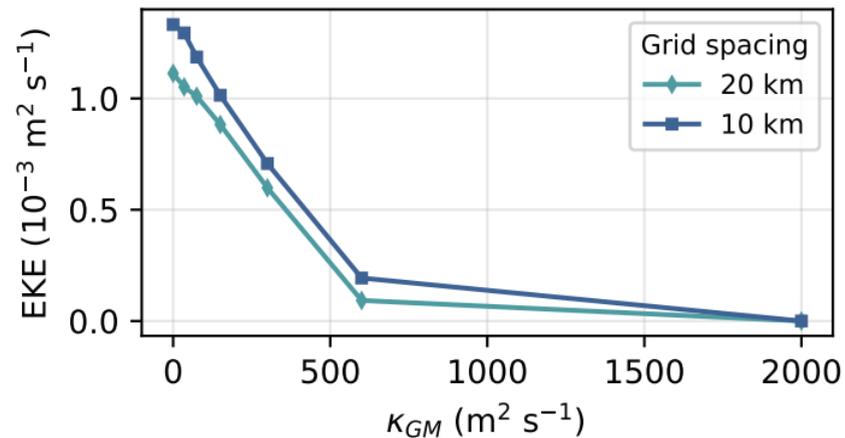
a. Gyre (20 km)



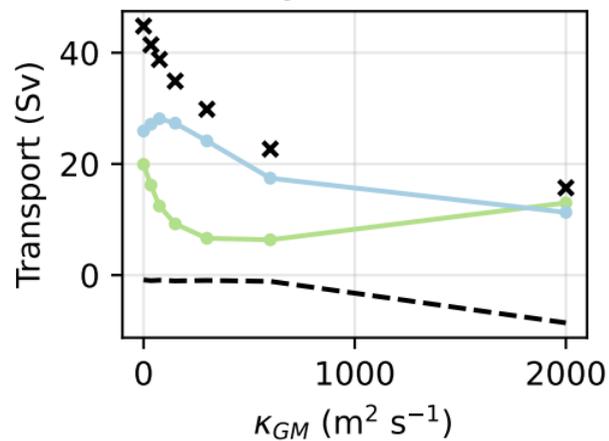
b. ACC (20 km)



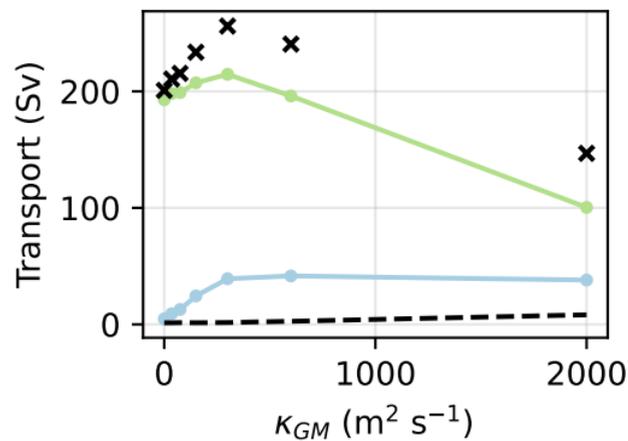
c. Mean EKE



d. Gyre (10 km)



e. ACC (10 km)



f. Gyre

