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## The contribution of submesoscale over mesoscale eddy iron transport in the open Southern Ocean

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1 **The contribution of submesoscale over mesoscale eddy**  
2 **iron transport in the open Southern Ocean**

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9 **Key Points:**

- 10 • A state of the art simulation at submesoscale permitting resolution coupled to a  
11 full biogeochemical model.
- 12 • Mesoscale vertical transport is first-order importance for iron supply from the in-  
13 terior to the surface.
- 14 • Submesoscale eddy transport becomes comparable to mesoscale near the surface  
15 where mixed-layer instability is active.

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**Abstract**

In order to examine the roles of ocean dynamics in supplying iron, the limiting nutrient in the open Southern Ocean, to the surface where it can be effectively utilized for photosynthesis, we run a flat-bottom zonally re-entrant channel model configured to represent the Antarctic Circumpolar Current region and couple it to a full biogeochemical model. The model was forced with monthly varying physical and biogeochemical boundary conditions to incorporate seasonality. Much focus on previous studies on iron pathways in the open ocean region has been on mixed-layer entrainment and diapycnal fluxes of iron. The Southern Ocean, however, is a region with strong meso- and submeso-scale turbulence and we would expect eddy fluxes to transport tracers including iron. Spatial resolution is, therefore, chosen as the parameter to control the effect of eddy transport. We utilize cross spectral analysis and the generalized Omega equation to temporally and spatially decompose the vertical transport attributable to meso- and submeso-scale motions. Our results suggest that the mesoscale vertical fluxes provide a first-order pathway for transporting iron across the mixing-layer base where diapycnal mixing is weak and must be included in modelling the open-Southern-Ocean iron budget.

**Plain language summary**

Ocean currents at the surface on the spatial scales of 1-200 km are energetic due to heating by the sun and stirring by the winds. These currents contribute significantly to the climate system by transporting heat and carbon horizontally towards the poles and vertically into the deep ocean. By running a numerical simulation at very high spatial resolution, we show that these currents are also responsible in transporting iron from the ocean interior to the surface in the Southern Ocean where phytoplankton growth is limited by the lack of iron, a key nutrient for most living organisms on Earth. Our results highlight the importance of accurately representing the ocean currents and associated iron transport in order to understand the Southern Ocean ecosystem and its impact on the climate via photosynthesis, the process in which carbon dioxide is converted to organic carbon and oxygen is produced as a bi-product.

**1 Introduction**

Ocean turbulence on scales of roughly 1-200 km is characterized by vigorous eddies, fronts, filaments, and other structures which collectively make an important con-

47 tribution to material transport. These motions can be categorized as either mesoscale  
48 or submesoscale depending on their degree of geostrophic balance (McWilliams, 2016).  
49 In the context of climate, it is well established that mesoscale transports are of first-order  
50 importance to the global ocean heat budget, especially for the vertical transport of heat  
51 (Wolfe et al., 2008; Griffies et al., 2015; Liang et al., 2015), and can drive global-scale  
52 variations in ocean heat content (Liang et al., 2017; Busecke & Abernathey, 2019). In  
53 the Southern Ocean, mesoscale eddies counteract the wind-driven circulation, playing  
54 a key role in the climate sensitivity of the global overturning circulation (Farneti et al.,  
55 2010; Zika et al., 2013; Gent, 2016).

56 As observations and simulations resolve smaller and smaller scales, the role of sub-  
57 mesoscale in determining large-scale ocean stratification has become a topic of interest.  
58 Due to the geostrophic inverse energy cascade (Kraichnan, 1967; Charney, 1971), mod-  
59 els that resolve submesoscales tend to have more energetic mesoscales than models with  
60 coarser resolution (Capet et al., 2008a, 2008b, 2008c). For instance, Lévy et al. (2010)  
61 showed that resolving the submesoscales in a basin-scale general circulation model (GCM)  
62 led to significant changes in circulation and stratification compared to lower-resolution  
63 runs. A coarse-graining analysis of the transports in their model revealed, however, that  
64 the impact of higher resolution on tracer transport was mostly due to more energetic *mesoscale*  
65 and large-scale flow, rather than the direct contribution of submesoscales even within  
66 the mixed layer (Lévy, Resplandy, et al., 2012, Fig. 7 in their paper). In contrast, a re-  
67 cent study using outputs from global submesoscale permitting general circulation model  
68 (GCM) argued that the near-surface vertical heat flux associated with submesoscale tur-  
69 bulence could dominate over the mesoscale on a global scale (Su et al., 2018). The field  
70 has evidently not reached a consensus on the relative contributions of the submesoscale  
71 vs. mesoscale tracer transport.

72 Similar ambiguities regarding the role of the submesoscale in ocean ecosystems ex-  
73 ist in the biogeochemical literature. Submesoscale features are apparent in surface chloro-  
74 phyll observations from high-resolution remote sensing data (Mahadevan, 2016; Lee &  
75 Kim, 2018). Recent studies have argued submesoscale vertical transport to be an impor-  
76 tant pathway in supplying nutrients to the euphotic layer (e.g. Lévy, Ferrari, et al., 2012;  
77 Levy & Martin, 2013; Mahadevan, 2016; Lévy et al., 2018; Zhang et al., 2019). Precisely  
78 isolating the relative contribution of this mechanism, however, has remained challeng-  
79 ing since as was noted earlier, submesoscale physics can feed back onto the large-scale

80 state. Lévy, Iovino, et al. (2012) coined a useful framework for thinking about this prob-  
81 lem: in analyzing simulations of increasing resolution, *local effects* are those impacts di-  
82 rectly attributable to the newly resolved small scales, while *remote effects* result from  
83 changes to the larger-scale flow which occur at higher resolution. In a simulation of a  
84 mid-latitude gyre system, Lévy, Iovino, et al. (2012) found that remote effects won out;  
85 as resolution was increased, phytoplankton abundance decreased, despite the increase  
86 in submesoscale nutrient flux.

87 The local vs. remote question has important implications for eddy parameteriza-  
88 tion in coarse-resolution models. If the submesoscale *local* effect proves to be significant,  
89 we would either need to resolve the submesoscale dynamics in such models or parametrize  
90 the submesoscale effect on nutrient transport in order to get the correct tracer estimates  
91 and predictions. On the other hand, if the main impact of submesoscales to enhance mesoscale  
92 transports via the inverse energy cascade (a *remote* effect), we may be able to rely on  
93 energy backscatter parametrizations which replicate the inverse energy cascade to en-  
94 ergize the mesoscale field without explicitly resolving the smallest scales (Jansen et al.,  
95 2015; Anstey & Zanna, 2017; Bolton & Zanna, 2019; Bachman, 2019).

96 In this paper, our aim is to quantify the impact of eddy fluxes on iron transport  
97 in the context of the Southern Ocean using idealized submesoscale-permitting simula-  
98 tions of varying resolution by disentangling carefully between *local* vs. *remote* effects.  
99 The Southern Ocean is one of the high-nutrient low-chlorophyll oceans (Field et al., 1998;  
100 Nolting et al., 1998; J. K. Moore & Abbott, 2000; Arrigo et al., 2008). Artificial fertil-  
101 ization experiments have shown that iron is predominantly the limiting nutrient in the  
102 open ocean region (Martin et al., 1990; De Baar et al., 1995; Coale et al., 2004; Aumont  
103 & Bopp, 2006). Recognizing the importance of iron in controlling Southern Ocean biomass  
104 and the associated biological carbon pump (Lévy et al., 2013), many past studies have  
105 attempted to quantify the iron supply, focusing primarily on boundary processes (i.e. ae-  
106 olian dust, and deposition by glacial melt, bathymetry and hydrothermal vents; Boyd  
107 & Ellwood, 2010; Boyd et al., 2012; Nishioka et al., 2013; Wadley et al., 2014). There  
108 has, however, been comparatively limited investigation into how ocean dynamics, specif-  
109 ically eddies and fronts, transport iron from the ocean interior (where it is high in con-  
110 centration) to the surface layer. Conventional focus has been on coastal processes (McGillicuddy  
111 et al., 2015; Mack et al., 2017; Jiang et al., 2019), mixed-layer entrainment of iron due  
112 to wintertime cooling (Tagliabue et al., 2014; Llorc et al., 2015, 2019) and storms (Carranza

113 & Gille, 2015; Nicholson et al., 2016), mesoscale isopycnal heaving (Swart et al., 2015;  
114 Song et al., 2016, 2018), and lateral stirring (d’Ovidio et al., 2015; Ardyna et al., 2017,  
115 2019). To the best of our knowledge, Rosso et al. (2014, 2016) are the only studies that  
116 examine the effect of submesoscale dynamics on iron supply to the surface ocean in the  
117 open-ocean region of Southern Ocean, but with the geographical coverage limited to the  
118 Kerguelen Plateau region and no consideration of the seasonal cycle.

119 While it would be ideal to quantify eddy iron transport using in-situ observations  
120 such as the Argo floats and Geotraces sections (Tagliabue et al., 2014), they lack the spa-  
121 tial and temporal resolution to sufficiently sample mesoscale and submesoscale features  
122 (Llort et al., 2018). Satellite observations on the other hand, while having good surface  
123 spatial coverage, cannot directly reveal vertical transport. We, therefore, turn to numeri-  
124 cal simulations and tackle the local vs. remote question by temporally and spatially de-  
125 composing the eddy *iron* fluxes. We accomplish this by applying cross spectral analy-  
126 sis, which allows us to examine the spatial and temporal scales of eddy transport, and  
127 the generalized Omega equation (Giordani & Planton, 2000). From the spectral perspec-  
128 tive, mesoscale fluxes have larger spatial and longer timescales than the submesoscale.  
129 The Omega equation, in contrast, provides a dynamics-based decomposition, decompos-  
130 ing the eddy transport into a mesoscale component in balance with the ageostrophic hor-  
131 izontal flow and a submesoscale component associated with higher Rossby numbers  
132 (McWilliams, 2016; McWilliams et al., 2019; Chereskin et al., 2019).

133 In this study, we adopt the flat-bottom zonally re-entrant channel framework de-  
134 veloped by Abernathey et al. (2011) and use spatial resolution as the parameter to mod-  
135 ulate the eddy effects; we run four cases ranging from mesoscale to submesoscale per-  
136 mitting resolutions: 20 km with and without eddy parametrization, 5 km and 2 km res-  
137 olution. Although the configuration is a rather strong idealization, it has been success-  
138 fully employed to investigate tracer transport in the Southern Ocean (Abernathey et al.,  
139 2013; Abernathey & Ferreira, 2015). The channel-like idealization is deliberate, as it lets  
140 us keep essentially the same mean flow and large-scale nutrient distribution across all  
141 simulations. This mitigates the confounding impacts of resolution on the basin-scale cir-  
142 culation reported in the North Atlantic study of Lévy, Iovino, et al. (2012); in their case,  
143 primary production decreased with increased resolution largely due to a change in the  
144 mean circulation, namely the Gulf Stream separation which has been shown to be sen-  
145 sitive to submesoscale boundary layer processes (Renault et al., 2016; Schoonover et al.,

146 2017). In contrast, the main potential remote effect of resolution in our simulations is  
 147 to energize the mesoscale. This setup therefore provides an ideal testbed to quantify the  
 148 impacts of mesoscale vs. submesoscale transport on phytoplankton ecology. In this study,  
 149 our focus is on the the resolution- and scale-dependence of physics and the physical drivers  
 150 for eddy iron transport. A deeper examination of the ecosystem dynamics and overall  
 151 controls on primary production is left to a companion paper in preparation (hereon U19).

152 Our paper is organized as follows: We describe the experimental setup briefly in  
 153 the next section, and the physical results are shown in section 3. Detailed analysis of sea-  
 154 sonal dynamics in the context of baroclinic instability and frontogenesis is found in sec-  
 155 tion 3.1, and the generalized Omega equation we use to decompose meso- and submeso-  
 156 scale motion in section 3.2. We show the biogeochemical results in section 4 with the em-  
 157 phasis on eddy vertical iron transport (Section 4.1), which is compared and contrasted  
 158 with the vertical eddy buoyancy flux. Conclusions are given in section 5.

## 159 2 Model description

### 160 2.1 Physical setup

161 The model setup is similar to Balwada et al. (2018) without a topographic ridge,  
 162 using the hydrostatic configuration of the Massachusetts Institute of Technology gen-  
 163 eral circulation model (MITgcm; Marshall et al., 1997). The channel domain ( $L_x = 1000$  km  
 164  $\times L_y = 2000$  km  $\times H = 2985$  m) is flat bottom and zonally re-entrant on a  $\beta$ -plane  
 165 centered around 49S ( $f_0 = -1.1 \times 10^{-4}$  s $^{-1}$ ,  $\beta = 1.4 \times 10^{-11}$  m $^{-1}$  s $^{-1}$ ). The horizon-  
 166 tal grids are on a Cartesian coordinate. The runs with 20 km resolution have 40 verti-  
 167 cal levels with 10 m near the surface. The 5 km and 2 km runs have 76 vertical levels  
 168 with 1 m near the surface. Monthly varying sea-surface temperature (SST) relaxation  
 169 and zonal wind stress are applied at the surface; SST increases from 0°C to 8°C from  
 170 south to north, and the zonal cosine-squared-shaped wind stress takes its maximum am-  
 171 plitudes between 0.1-0.2 N m $^{-2}$  at the center of the meridional extent and is tapered to  
 172 zero at the northern and southern 50 km extent of the domain. The exact formulations  
 173 are given in Appendix A1. The Leith-scheme horizontal (Pearson et al., 2017) and ver-  
 174 tical viscosity values of  $A_h = 2.15$  m $^2$  s $^{-1}$  and  $A_v = 5.6614 \times 10^{-4}$  m $^2$  s $^{-1}$  are used.  
 175 We apply no-slip boundary conditions at the channel walls and bottom with the latter

176 having a quadratic drag,  $C_d = 2.1 \times 10^{-3}$ . Other parameter values are given in Ta-  
 177 ble A1.

178 We run four cases: 20, 5 and 2 km horizontal resolution with 20 km run having the  
 179 case of i) no eddy parametrization, and ii) mixed-layer instability (MLI; Boccaletti et  
 180 al., 2007; Fox-Kemper et al., 2011) parametrization turned on. The 20 km run was spun  
 181 up from a state of rest for 200 years and each run with higher resolution was subsequently  
 182 spun up from the climatology of the lower resolution run until the domain-averaged tem-  
 183 perature field reached a statistical equilibrium. Mixed-layer instability is a type of baro-  
 184 clinic instability that feeds off the available potential energy (APE) within the mixed layer  
 185 and the parametrization restratifies the mixing layer (ML) due to the instability if oth-  
 186 erwise resolved (Fox-Kemper et al., 2008; Fox-Kemper & Ferrari, 2008; Fox-Kemper et  
 187 al., 2011). We show in U19 that the MLI parametrization, intended for restratifying the  
 188 ML and not for eddy tracer transport, does not enhance vertical tracer transport com-  
 189 pared to the non-parametrized run. Considering that 20 km resolution is mesoscale per-  
 190 mitting and not resolving (*grey zone*), one may also consider in addition to MLI, turn-  
 191 ing on the Gent-McWilliams’ eddy-induced velocity parametrization (GM; Gent & McWilliams,  
 192 1990; McDougall & McIntosh, 2001) developed to extract APE due to otherwise resolved  
 193 mesoscale baroclinic instability. Hallberg (2013), using a two-layer Phillips-type isopy-  
 194 cnal model, however showed that the GM parametrization for models in the grey zone  
 195 acts to suppress the resolved eddies rather than replicate their effects. The suppression  
 196 of eddy effects was effective in our case as well; turning on GM resulted in steepening  
 197 of the isopycnals rather than slumping them (not shown). We, therefore, do not have  
 198 a case with GM turned on. Further details of the physical boundary conditions and pa-  
 199 rameters are given in A1 and Table A1.

The Rossby deformation radius ( $R_d$ ) in all resolution runs at the center of the do-  
 main is roughly 14 km. The radius was obtained by solving the Sturm-Liouville eigen-  
 value problem

$$\frac{d}{dz} \left( \frac{f^2}{N^2} \frac{d\xi}{dz} \right) = -\kappa^2 \xi \quad (2.1.1)$$

200 where  $\kappa$  corresponds to the Rossby deformation wavenumber and radius is the inverse  
 201 wavenumber ( $R_d = \kappa^{-1}$ ).  $\xi(z)$  is the vertical mode associated with each wavenumber.  
 202 Equation 2.1.1 is derived from the linearized quasi-geostrophic potential vorticity equa-  
 203 tion around a state of rest and prescribing a plane-wave solution (Vallis, 2017, Sections 5.8.2, 9.4.3



204 in their book). We used the seasonal-zonal mean of stratification as the background state.  
 205  $R_d$  does not change seasonally, since the interior stratification does not have a strong sea-  
 206 sonal cycle. The spatial resolution of 2 km is roughly a decade smaller than  $R_d$  allow-  
 207 ing for partially resolved submesoscale dynamics (Lévy et al., 2018); relative vorticity  
 208 ( $\zeta$ ) reaches up to three times the local Coriolis parameter indicated by the Rossby num-  
 209 ber ( $Ro = \frac{\zeta}{f}$  where  $f$  is the Coriolis parameter; Fig. 1d). Idealized models serve as a  
 210 valuable tool to investigate the physical drivers of seasonality in (sub)mesoscale turbu-  
 211 lence (Qiu et al., 2014; Brannigan et al., 2015). To the best of our knowledge, our study  
 212 is the first idealized modelling study to compare the MLI parametrization with subme-  
 213 soscale permitting simulations that cover the temporal and spatial range allowing for eddy-  
 214 eddy and eddy-mean flow interaction under seasonal forcings, for which we show the re-  
 215 sults in section 3.

## 216 2.2 Biogeochemical setup

217 We couple the seasonally resolving physical simulation to the ecosystem model of  
 218 Gloege et al. (2017), which simplifies the Darwin biogeochemical model (Follows et al.,  
 219 2007) to the complexity of the two species ecosystem described in Dutkiewicz et al. (2009);  
 220 there are two phytoplankton (diatoms and small phytoplankton) and two zooplankton  
 221 functional groups respectively. The general physiology of each phytoplankton species is  
 222 that diatoms have a faster maximum growth rate but favor conditions with high nutri-  
 223 ent concentrations, while the small phytoplankton are more resilient in low nutrient and  
 224 light environments but have a slower maximum growth rate (Fig. A2). The model con-  
 225 siders the full biogeochemical cycle of phosphate, nitrate, silicate, carbon, oxygen and,  
 226 for the interest of our study, iron, with 31 distinct prognostic tracers advected and dif-  
 227 fused by the flow.

228 Domain-wide nutrient supply ( $PO_4$ ,  $NO_3$ , Fe and  $SiO_2$ ) is accomplished via a sponge  
 229 layer at the northern 100 km extent of the domain. These nutrients then freely evolve  
 230 in the interior following the circulation and biogeochemical cycles. The relaxation pro-  
 231 files for  $PO_4$ ,  $NO_3$  and  $SiO_2$  were taken from the World Ocean Atlas (WOA) at 45S and  
 232 then interpolated onto our model vertical grid. We use the monthly climatological prod-  
 233 ucts down to 500 m where monthly data is available and append the annual climatol-  
 234 ogy below. Monthly iron profiles were taken from the Biological Southern Ocean State  
 235 Estimate (BSOSE; Verdy & Mazloff, 2017) as the Geotraces dataset (Tagliabue et al.,

236 2012, 2014) did not have sufficient temporal and spatial resolution. In an effort to com-  
 237 pensate for the lack of dust, glacial and bathymetric sources, we chose 50S of BSOSE,  
 238 which had higher concentrations than at 45S, but details of the relaxation profiles ul-  
 239 timately did not make a difference in surface concentrations as the spun up interior iron  
 240 concentration was rather insensitive to the details of the relaxation profile (not shown).  
 241 The zonal mean of each product for all four nutrients were taken across 50-150E in the  
 242 Kerguelen Plateau region. Photosynthetically available radiation (PAR) is prescribed  
 243 at the surface as a meridional linear fit to the monthly-zonal mean of SeaWiFS prod-  
 244 uct between latitudes of 45-60S taking its minimum (maximum) in June (December; Fig. A2a-  
 245 c).

246 In order to isolate the effects of iron supply by ocean dynamics from depth in the  
 247 open ocean region, we did not include dust deposition, glacial melt and bathymetric sources.  
 248 Our ecosystem is consequently iron limited year round (Fig. A3) and pelagic commu-  
 249 nity transition does not occur; diatoms dominate the community year round whereas in  
 250 the real Southern Ocean, silicate limitation likely comes into play (J. K. Moore et al.,  
 251 2004; C. Moore et al., 2013; Carranza & Gille, 2015). Due to this dominance by diatoms,  
 252 our springtime vertically integrated phytoplankton carbon biomass reaches its apex (timing  
 253 of  $\langle C_p \rangle$  maximum where  $\langle \cdot \rangle = \int \cdot dz$ ; Behrenfeld, 2010) in early November, roughly 1-  
 254 2 months earlier than estimates from biogeochemical (BGC) Argo floats in the South-  
 255 ern Ocean (Appendix B, Fig. B1; Uchida et al., submitted).

256 Although we believe that it would be possible to further tune the biogeochemical  
 257 model, parameters tuned for lower resolution runs were not directly applicable to im-  
 258 prove the ecosystem in higher resolution runs due to changes in surface iron concentra-  
 259 tion; vertical eddy iron transport increased and the spring bloom tended to occur ear-  
 260 lier in the year with resolution for the same biogeochemical parameters. Since 2-km res-  
 261 olution coupled to a full biogeochemical model is state of the art in terms of resolution,  
 262 we decreased the growth rates from those used in Dutkiewicz et al. (2009) within the ac-  
 263 ceptable parameter range of previous studies (e.g. Dutkiewicz et al., 2009; Bennington  
 264 et al., 2009; Llort et al., 2015; Gloege et al., 2017) as a one-shot attempt to achieve a  
 265 reasonable spring bloom. The biogeochemical parameter values were then kept identi-  
 266 cal for all coarser runs with the maximum growth rate for diatoms and small phytoplank-  
 267 ton being  $0.81 (= 1/1.24) \text{ days}^{-1}$  and  $0.56 (= 1/1.8) \text{ days}^{-1}$  respectively (Table A2).  
 268 Considering the agreement in magnitude of the seasonal cycle and timing of bloom on-

269 set occurring around July (Fig. B1), we argue that our model, although idealized, serves  
 270 as a valuable tool in quantifying the eddy transport of iron and interaction of physics  
 271 and biogeochemistry at submesoscales, which is the focus of this study.

272 We spun up the biogeochemistry for another five years after the dynamics had spun  
 273 up until the domain averaged iron concentration reached a statistical equilibrium for the  
 274 20 and 5 km run. For the 2 km run, we spun up the biogeochemistry for the latter 2.5 years  
 275 of the total spin up simultaneously with the dynamics at which the iron concentration  
 276 over the meridional extent we analyze ( $y \in [600, 1400]$  km) reached statistical equilib-  
 277 rium. Further details of the biogeochemical boundary conditions and parameter values  
 278 are given in Appendix A2 and Table A2.

### 279 **3 Physical results**

280 We start by showing a snapshot of the local Ro in the top 300 m for the 2 km run  
 281 in September 15 representative of austral winter (Fig. 1a), and February 15 as represen-  
 282 tative of austral summer (Fig. 1b). The skewness and seasonal difference of the prob-  
 283 ability density function (PDF) of Ro increases with resolution; the 20 km shows very lit-  
 284 tle seasonality with the winter and summer PDFs overlying on top of each other. Well-  
 285 formed mesoscale eddies develop year round, and it is clear that winter time has higher  
 286 Ro values and finer features than in summer (Fig. 1d).

287 The effect of seasonal forcing in temperature and wind stress can be seen in SST  
 288 and *mixing*-layer depth (MLD) as daily-spatial means in Fig. 2. The resulting SST takes  
 289 its maximum in February and minimum in September (Fig. 2a) consistent with the sea-  
 290 sonal cycle in the ACC region of BSOSE (not shown). The MLD is the depth over which  
 291 isotropic mixing is active due to surface wind stress and diabatic forcing; here we define  
 292 this highly variable depth as the zonal 99th percentile of the daily-averaged K-profile parametriza-  
 293 tion (KPP; Large et al., 1994) boundary layer. We argue that it is the *mixing* layer (ML)  
 294 and not *mixed* layer that is relevant for tracer subduction/obduction as ML is the layer  
 295 over which mixing is active (Balwada et al., 2018). The MLD averaged over the merid-  
 296 ional extent of  $y \in [600, 1400]$  km (in order to avoid the channel wall effects) is the deep-  
 297 est during September and shallowest in January (Fig. 2b), slightly out of phase with the  
 298 seasonal cycle of SST during summer. This shows the the ML variability is buoyancy  
 299 driven and not the winds which take a bi-annual structure (Appendix A1).

300 We see that as the spatial resolution increases, the ML becomes shallower, which  
 301 is expected from mixed-layer instability (MLI; explained in further detail in the section  
 302 below) effectively restratifying the ML. We attribute the 20 km runs having shallower  
 303 MLD than the 2 and 5 km run over the summer (January, February, March; JFM) to  
 304 the difference in vertical resolution. Comparing the 20 km and 20 km MLI run, we see  
 305 the parametrization effectively restratifies the ML; the wintertime MLD maximum shoals  
 306 roughly 100 m. The SSTs for the two runs, however, are essentially identical, implying  
 307 that the parallel increase in SST we see with resolution is due to the mesoscale restrat-  
 308 ification of the interior rather than MLI.

### 309 **3.1 Mixed-layer instability as the seasonal driver of submesoscale tur-** 310 **bulence**

311 It is common to quantify the temporal variability, seasonality in our case, by ex-  
 312 amining the kinetic energy (KE) of the system (Sasaki et al., 2014; Callies et al., 2015;  
 313 Rocha et al., 2016; Uchida et al., 2017). We remove the zero-th zonal wavenumber com-  
 314 ponent (i.e. zonal mean) as the deviation in snapshot outputs every 15 days from the  
 315 mean, viz.  $\mathbf{u}' = \mathbf{u} - \bar{\mathbf{u}}$  where the overbar denotes the seasonal and zonal mean. The  
 316 15-day interval was chosen as the time scale at which the autocorrelation of daily-averaged  
 317 horizontal velocity anomaly ( $u'$ ) at the center of the domain crossed zero (not shown);  
 318 we treat each anomaly field ( $\mathbf{u}'$ ) as an individual realization of the turbulence process  
 319 in time. We take the zonal Fourier transform of  $\mathbf{u}'$  and temporally average them to con-  
 320 struct seasonal-mean spectra  $\overline{|\hat{\mathbf{u}}'|^2}$  where  $(\hat{\cdot}) = \int(\cdot)e^{ikx} dx$  is the zonal Fourier trans-  
 321 form). Since our model is a re-entrant channel, all of our wavenumber spectra were taken  
 322 in the zonal direction, without any tapering applied, using the Python package `xrft` ([https://](https://xrft.readthedocs.io/en/latest/)  
 323 `xrft.readthedocs.io/en/latest/`) and then averaged over the meridional extent of  
 324  $y \in [600, 1400]$  km. As a reference to our zonal-mean view of the Southern Ocean, the  
 325 climatological zonal-wavenumber spectra of AVISO geostrophic KE is shown as well us-  
 326 ing daily-averaged velocity fields sampled every 15 days. The AVISO zonal-wavenumber  
 327 spectra were taken at latitudes between 50-60S wrapping zonally around the globe and  
 328 then averaged meridionally assuming a Cartesian plane between those latitudes. The zonal  
 329 wavenumber KE spectra (Fig. 3) shows three things: i) the mesoscales ( $O(50$  km)) are  
 330 more energetic for higher resolution runs similar to Capet et al. (2008a, Fig. 6 in their  
 331 paper), ii) wintertime has higher KE at scales below 25 km, and iii) large scales (wavenum-

332 bers corresponding to scales above  $O(100 \text{ km})$ ) have the same order of magnitude as AVISO  
 333 observations in the SO. Integrating the KE spectra from the 2 km run over scales be-  
 334 low 25 km and plotting it against MLD shows that their seasonality is in phase, i.e. high  
 335 KE with deep ML and visa versa (Fig. 3b).

Following Uchida et al. (2017), we quantify the mechanism for the surface KE sea-  
 sonality through baroclinic instability and frontogenesis. Frontogenesis is a process in  
 which mesoscale stirring generates submesoscale filaments by bringing buoyancy contrasts  
 closer, setting up localized sources for instabilities (McWilliams, 2016) and can be quan-  
 tified by the frontogenesis function defined as

$$F_s = \mathbf{Q}_s \cdot \nabla_h b \quad (3.1.1)$$

336 where  $\mathbf{Q}_s = -\left(\frac{\partial u}{\partial x} \frac{\partial b}{\partial x} + \frac{\partial v}{\partial x} \frac{\partial b}{\partial y} + \frac{\partial w}{\partial x} \frac{\partial b}{\partial z}, \frac{\partial u}{\partial y} \frac{\partial b}{\partial x} + \frac{\partial v}{\partial y} \frac{\partial b}{\partial y} + \frac{\partial w}{\partial y} \frac{\partial b}{\partial z}\right)$  indicating whether the  
 337 flow field increases or decreases the buoyancy gradients (Hoskins, 1982; Capet et al., 2008b;  
 338 Brannigan et al., 2015) and was calculated using 15-daily snapshot outputs. Buoyancy  
 339 was defined using temperature only as we use a linear equation of state with no salin-  
 340 ity ( $b = \alpha g \theta$ ).

341 Figure 4 shows a time-depth Hovmöller diagram of baroclinic available potential  
 342 energy (APE) conversion rate ( $\overline{w'b'}$ ), frontogenesis function normalized by the magni-  
 343 tude of horizontal buoyancy gradient ( $\overline{F_s/|\nabla_h b|}$ ), and the root-mean square of vertical  
 344 velocity for the 5 and 2 km run (we do not show the 20 km run as the amplitudes were  
 345 orders of magnitude smaller). The amplitude of each increases with resolution, which  
 346 is expected from better resolved fronts and mixed-layer instability (MLI; Boccaletti et  
 347 al., 2007) with higher resolution. We see a strong seasonality of  $\overline{w'b'}$ , an indicator of MLI  
 348 within the ML. In a spatial mean sense,  $\overline{w'b'}$  and  $\overline{F_s/|\nabla_h b|}$  are both positive year round  
 349 with the latter more surface intensified. This implies that frontogenesis always acts to  
 350 strengthen the buoyancy fronts on which MLI feeds off, converting APE to KE. Fron-  
 351 togenesis acting to strengthen the fronts over the summer when ML is shallow is con-  
 352 sistent with the seasonal cycle of submesoscale turbulence found by Brannigan et al. (2015),  
 353 but we also find large positive values over the winter, which may be due to different sur-  
 354 face forcing conditions and/or eddy-mean flow interaction; Brannigan's domain size was  
 355 too small to allow for any eddy-eddy and eddy-mean flow interactions. The phasing be-  
 356 comes clearer when we take the depth average over the top 100 m of  $\overline{w'b'}$  and  $\overline{F_s/|\nabla_h b|}$   
 357 with the two in phase with KE (Fig. 3b,c). This is consistent with Uchida et al. (2017)

358 who, using outputs from a ocean-atmosphere fully coupled GCM, showed that even par-  
 359 tially resolved MLI can modulate seasonality in surface KE.

Based on the the increase in surface KE with resolution (Fig. 3), we hypothesize that MLI in the submesoscale range acts as an energy source for the inverse energy cascade to energize the mesoscales ( $O(50\text{ km})$ ; Charney, 1971; Arbic et al., 2013; Qiu et al., 2014; Barkan et al., 2015; Callies et al., 2016), and forward cascade, which is here represented by the numerical Leith scheme dissipation (Pearson et al., 2017). We quantify this by taking the seasonal and meridional mean of the zonal wavenumber cross spectra of  $w'$  and  $b'$ , and the KE spectral flux

$$\Pi(k) = - \int_{\kappa > k} \mathcal{R}[\hat{\mathbf{u}} \cdot (\mathbf{u} \cdot \hat{\nabla}_h \mathbf{u})^*] d\kappa \quad (3.1.2)$$

360 where  $\mathbf{u}$  is the 15-daily snapshot output of total horizontal velocity. The former allows  
 361 us to quantify the spatial scales at which APE conversion to KE due to baroclinic in-  
 362 stability is active and the latter the direction of KE cascade; in the framework of geostrophic  
 363 turbulence we would expect KE to cascade upscale ( $\Pi < 0$ ; Charney, 1971; Arbic et al.,  
 364 2013). As was noted earlier for the zonal-wavenumber KE spectra, the zonal re-entrant  
 365 configuration allows us circumvent introducing artificial wavenumber modes and spuri-  
 366 ous errors by tapering, which Aluie et al. (2018) showed by comparing tapered spectral  
 367 fluxes to their energetically consistent coarse-graining method. In other words, our spec-  
 368 tral flux is exact in the zonal dimension.

369 The APE conversion rate in the surface 200 m takes its maxima at scales  $O(30\text{ km})$   
 370 for the 2-km run, which coincides with where seasonality in surface KE is apparent (Figs. 3a, 5a,b).  
 371 Associated with the surface maxima of APE conversion, there is a change in sign in the  
 372 spectral flux around  $O(30\text{ km})$  particularly during winter at scales reaching into higher  
 373 wavenumbers than in summer and the magnitude increases with resolution (Fig. 5e-h),  
 374 consistent with the findings by Capet et al. (2008c); Sasaki et al. (2014). The 5-km run  
 375 has a much lower APE conversion rate and consequently a weaker KE spectral flux (Fig. 5c,d,g,h).  
 376 This confirms that MLI acts as an energy source for the inverse KE cascade ( $\Pi < 0$ )  
 377 particularly during winter with positive values of  $w'b'$  (Fig. 5a,b) at scales around the  
 378 Rossby deformation radius, resulting in better agreement between AVISO and higher res-  
 379 olution runs in the mesoscale range (Fig. 3a). The forward cascade ( $\Pi > 0$ ) at the small-  
 380 est scales is due to the Leith-scheme viscosity (Bachman et al., 2017). The large signal  
 381 of inverse KE cascade at scales above  $O(100\text{ km})$  is likely coming from the deep mesoscale

382 baroclinic instability (Fig. 5a). We argue in section 4 that accurate representation of mesoscale  
 383 dynamics is crucial for modulating the eddy iron transport.

### 384 **3.2 Decomposing the (sub)mesoscales using the Omega equation**

385 With the advent of submesoscale permitting GCMs, the relative importance of sub-  
 386 mesoscale heat flux over mesoscale has been an active topic of research; Su et al. (2018)  
 387 argued that on a global scale, submesoscale vertical heat flux could dominate over the  
 388 mesoscale. Although the zonal-wavenumber cross spectra is already telling that verti-  
 389 cal buoyancy flux associated with submesoscale turbulence is significant in our case as  
 390 well (Fig. 5a,b), we can further dynamically decompose the transport into its balanced  
 391 and unbalanced component. Instead of an ad-hoc temporal or spatial filter commonly  
 392 applied to decompose the fluxes (Uchida et al., 2017; Su et al., 2018), here we use the  
 393 generalized Omega equation (Giordani & Planton, 2000) to diagnose the vertical flow  
 394 field in balance with forcing (Molemaker et al., 2010).

The Omega equation is purely diagnostic in a sense that it includes no terms with  
 a time derivative and takes the form

$$N^2 \nabla_{\text{h}} w_{\text{b}} + f_0^2 \frac{\partial^2 w_{\text{b}}}{\partial z^2} = \beta \frac{\partial b}{\partial x} + \nabla_{\text{h}} \cdot \mathbf{Q}(\mathbf{u}, b, \Phi) \quad (3.2.1)$$

395 where  $\mathbf{Q}$  is a function of the instantaneous horizontal velocities, buoyancy and pressure  
 396 for which the exact form will be given in Appendix C. We define the inverted velocity  
 397 from eqn. 3.2.1 as the balanced motion ( $w_{\text{b}}$ ) and residual from the total vertical veloc-  
 398 ity as the unbalanced motion ( $w_{\text{ub}} = w - w_{\text{b}}$ ) and here, we simply show a snapshot  
 399 example of the inversion at the depth of  $z = -211$  m (Fig. 6). To first order, the Omega  
 400 equation behaves as a low-pass filter, as we can see from Fig. 6a-c that  $w_{\text{b}}$  captures the  
 401 large-scale features in balance with the ageostrophic horizontal velocities (Giordani &  
 402 Planton, 2000) and  $w_{\text{ub}}$  the fronts with superposition of waves. It is no surprise that the  
 403 contribution of  $w_{\text{ub}}$  is large near the surface and bottom where MLI and boundary layer  
 404 processes are active, while  $w_{\text{b}}$  captures most of the variance in the interior (Fig. 6d). As  
 405 a reference, we also show the root-mean square profile of vertical velocity inverted from  
 406 the quasi-geostrophic Omega equation ( $w_{\text{qg}}$ ; Hoskins et al., 1978). We see that includ-  
 407 ing higher-order ageostrophic terms captures the variance of total vertical velocity to a  
 408 better extent (compare  $w_{\text{b}}$  and  $w_{\text{qg}}$ ).

409 In order to show the scale separation, it is useful to take the wavenumber power  
 410 spectra of each component. From Fig. 5a,b, it is apparent that wintertime has higher  
 411 submesoscale activity so we will focus only on wintertime for the dynamical decompo-  
 412 sition. Figure 7a-c shows that the balanced motion has higher power at larger scales than  
 413 the unbalanced motion, with the latter being surface intensified. The decomposition in  
 414 power spectra is not exact, as there is correlation between the balanced and unbalanced  
 415 motion, i.e.  $\hat{w}^2 = \hat{w}_b^2 + \hat{w}_{ub}^2 + 2\hat{w}_b\hat{w}_{ub}^*$  but the power spectra of each component is a  
 416 good qualitative indicator of the scale separation.

417 For the cross spectra of vertical velocity and buoyancy, however, the decomposi-  
 418 tion is exact, i.e.  $\hat{w}\hat{b}^* = \hat{w}_b\hat{b}^* + \hat{w}_{ub}\hat{b}^*$ . The total vertical velocity ( $w$ ) used here for  
 419 the decomposition is slightly different from  $w'$  in Fig. 5a,b in that the seasonal-zonal mean  
 420 was not subtracted out as we use the total horizontal velocity, potential temperature and  
 421 pressure fields in inverting for  $w_b$  (eqn. 3.2.1); Figures 5a and 7d differ by the seasonal  
 422 mean component but we find the difference is negligible as apparent in comparing the  
 423 two figures. Due to the zonally re-entrant configuration, a temporal mean and zonal mean  
 424 are statistically equivalent, but the difference indicates that the seasonal mean in the one  
 425 year of data has a zonal structure. In other words,  $w_{ub}$  includes both the unbalanced and  
 426 seasonal mean motion with the latter likely causing the negative values at depths below  
 427 400 m at scales larger than  $O(100 \text{ km})$  in Fig. 7f counteracting the mesoscale eddies. The  
 428 vertical buoyancy flux associated with the unbalanced motion is more surface intensi-  
 429 fied than the balanced, consistent with our understanding of MLI (Fig. 7d-f). The sig-  
 430 nal at the very surface at scales above  $O(100 \text{ km})$  in  $\mathcal{R}[\hat{w}_{ub}\hat{b}^*]$  is an artifact likely due  
 431 to the Monotone Piecewise Cubic Interpolation scheme used to interpolate the  $w$  and  
 432  $b'$  fields onto the same monotonic vertical grid as  $w_b$  in order to diagnose  $w_{ub}$ . Looking  
 433 at the vertical profile of each component (Fig. 7g), there is a transition around  $z = -100 \text{ m}$   
 434 from the unbalanced motion to balanced motion being dominant in fluxing buoyancy.

#### 435 **4 Biogeochemical results**

The main goal of our study is to examine how eddy iron transport behaves in our  
 seasonally resolving and submesoscale permitting model coupled to a full biogeochem-  
 ical model. To this end, we solve the governing equation for iron online in MITgcm im-  
 plemented as part of the two-species biogeochemical model (Section 2.2; Dutkiewicz et



al., 2009)

$$\frac{\partial \text{Fe}}{\partial t} + \nabla \cdot (\mathbf{v}\text{Fe}) = \dot{\text{Fe}} + D_{\text{Fe}} \quad (4.0.1)$$

where the left-hand side is the familiar form of tendency and advection of passive tracers. The first term on the right-hand side ( $\dot{\text{Fe}}$ ) is the source/sink term which comes from primary production and  $D_{\text{Fe}}$  the diffusion. Since our domain is zonally re-entrant, it is natural to consider zonal-mean quantities, for which we show the zonal-seasonal mean iron budget:

$$\frac{\partial \overline{\text{Fe}}}{\partial t} = -\frac{\partial}{\partial y}(\overline{v\text{Fe}}) - \frac{\partial}{\partial z}(\overline{w\text{Fe}}) + \overline{\dot{\text{Fe}}} + \overline{D_{\text{Fe}}}. \quad (4.0.2)$$

437 We have no contribution from zonal advection due to the zonally re-entrant configura-  
 438 tion, viz.  $\overline{\partial_x(u\text{Fe})} = 0$ . In the following analysis, we only consider the meridional ex-  
 439 tent of  $y \in [600, 1400]$  km away from the north/south walls.

440 In order to put vertical eddy transport into perspective of the other terms in eqn. (4.0.2),  
 441 we calculate the zonal-seasonal mean iron budget for winter (July, August, September;  
 442 JAS) and summer (January, February, March; JFM) using daily averaged outputs (Fig. 8).  
 443 The eddy transport terms were obtained from daily averaged outputs by applying a Reynolds  
 444 decomposition, i.e.  $\overline{\mathbf{v}'\text{Fe}'} = \overline{\mathbf{v}\text{Fe}} - \overline{\mathbf{v}\text{Fe}}$ . It is clear that the vertical eddy transport ( $\overline{w'\text{Fe}'}$ ;  
 445 red dashed) is of first-order importance for both seasons, particularly during winter when  
 446 MLI is active, in the budget. The eddy transport reaches deeper into the water column  
 447 to bring up iron indicated by positive values ( $-\partial_z(\overline{w'\text{Fe}'}) > 0$ ) than the diffusive fluxes  
 448 ( $\partial_z \overline{F}_{\text{diff}}$ ; blue), in our case due to KPP mixing. The diffusive flux is convergent near the  
 449 surface, with the KPP mixing transporting iron down the vertical gradients actively gener-  
 450 erated by the biogeochemical sink at the surface and eddy iron supply from the interior  
 451 of iron. The net biogeochemical source/sink term ( $\overline{\dot{\text{Fe}}}$ ) is a net sink near the surface year  
 452 round due to primary production ( $\overline{\dot{\text{Fe}}_p}$ ) overwhelming the source by remineralization ( $\overline{\dot{\text{Fe}}_r}$ ;  
 453 green dashed). The contribution due to horizontal eddy transport ( $-\partial_y(\overline{v'\text{Fe}'})$ ; red dot-  
 454 ted) and mean advection ( $-\nabla \cdot (\overline{\mathbf{v}\text{Fe}})$ ; red solid) is small compared to the other terms  
 455 in our simulation.

#### 456 4.1 (Sub)mesoscale eddy iron transport

457 Given its dominant role in the budget, from here on we focus on vertical eddy iron  
 458 transport. Lévy et al. (2001), in the context of an oligotrophic ecosystem in a baroclin-  
 459 ically unstable jet, showed that nutrient supply increased with spatial resolution of their

460 model; vertical nutrient transport increased both along the submesoscale fronts and with  
 461 mesoscale vertical velocities energized via the inverse energy cascade. As a reminder, we  
 462 define the former as the *local* and latter as the *remote* effect. By definition, we would  
 463 expect mesoscale eddy transport to have longer timescales and larger spatial scales than  
 464 submesoscale transport. Our goal here is to quantify the spatial and temporal scales of  
 465 eddy iron transport and their relative contributions. We do so by taking the frequency  
 466 and zonal wavenumber ( $\omega$ - $k$ ) spectra (Fig. 9) of the eddy terms—defined as the devia-  
 467 tion in hourly snapshot outputs from the zonal and seasonal mean ( $Fe' = Fe - \overline{Fe}$ ). The  
 468  $\omega$ - $k$  spectra requires model outputs saved at high frequency for which we use hourly snap-  
 469 shot outputs, and buoyancy snapshot fields were only saved every 15 days. We, there-  
 470 fore, do not show the  $\omega$ - $k$  cross spectra for the buoyancy flux.

471 The  $\omega$ - $k$  power spectra of vertical velocity show differences between summer and  
 472 winter at scales associated with mesoscale turbulence ( $O(30\text{ km})$ ,  $O(\omega/f) \sim 0.2$ ) with  
 473 wintertime having higher power. There is also a signal of internal waves at super-inertial  
 474 frequencies ( $O(\omega/f) > 1$ ; Fig. 9a,b). When examining the cross spectra of vertical ve-  
 475 locity and iron, however, the signals at high frequency and wavenumber vanish for both  
 476 winter and summer (Fig. 9c,d). This implies that at  $z = -211\text{ m}$  depth which is right  
 477 below the MLD (Fig. 4d-f): i) eddy iron transport associated with mesoscale turbulence  
 478 associated with scales larger than the Rossby deformation radius dominates over sub-  
 479 mesoscale, and ii) waves associated with super-inertial frequencies contribute to no net  
 480 iron transport, consistent with the results by Balwada et al. (2018, Fig. 4 in their pa-  
 481 per) which found that internal waves are inefficient in transporting passive tracers. The  
 482 results were qualitatively similar at  $z = -180\text{ m}$  within the wintertime ML where al-  
 483 though there was an increase at larger wavenumbers, no increase at super-inertial fre-  
 484 quencies in the vertical iron transport (not shown). Comparing the cross spectra of the  
 485 2- to 5-km run (Fig. 9c-f), it is obvious that the vertical eddy transport is weaker in the  
 486 latter. We attribute this to the insufficient inverse energy cascade and resulting ener-  
 487 getically weak mesoscale field as was discussed in section 3.1.

488 Following our argument for vertical buoyancy fluxes, we also decompose the iron  
 489 fluxes using the Omega equation, i.e.  $\hat{w}Fe^* = \hat{w}_b\hat{F}e^* + \hat{w}_{ub}\hat{F}e^*$ . The scale separation is  
 490 again evident in Fig. 10a-c where the unbalanced transport has its maximum at smaller  
 491 scales. It is interesting to note that the vertical structure of buoyancy and iron trans-  
 492 port by the unbalanced motions are quite different. The eddy buoyancy flux is more sur-

493 face intensified (Fig. 7c), while the eddy iron flux is strongest between 100 m and 200 m  
 494 depth. One might question whether the large amplitude in iron transport at high wavenum-  
 495 bers in the unbalanced motion is due to submesoscale turbulence or internal waves (Fig. 10c).  
 496 The  $\omega$ - $k$  spectra (Fig. 9), however, show no indications of transport in the frequency range  
 497 of internal waves ( $\omega/f > 1$ ) so we can safely attribute iron transport due to the un-  
 498 balanced motion to submesoscale turbulence. Returning to the discussion between the  
 499 *local* and *remote* effect, Fig. 10d shows that, below the top 100 m (where the flux is weak),  
 500 the transport associated with balanced motion (remote) is larger than the unbalanced  
 501 component (local); however, unbalanced motions still contribute about 1/3 of the total  
 502 flux at 200 m depth.

## 503 5 Discussion and Conclusions

504 By running a seasonally resolving model at submesoscale permitting resolution con-  
 505 figured to represent the zonal-mean view of the Antarctic Circumpolar Current region,  
 506 our model partially resolves mixed-layer instabilities (MLI; Boccaletti et al., 2007) and  
 507 generates well formed mesoscale eddies and fronts (Fig. 1). The agreement between sur-  
 508 face KE in our model at scales larger than  $O(100 \text{ km})$  with KE estimates from satellite  
 509 altimetry observations improves with higher spatial resolution, as seen in the wavenum-  
 510 ber power spectra (Fig. 3a), likely as a result of a better-resolved inverse energy cascade  
 511 (Fig. 5; Capet et al., 2008a, 2008b, 2008c; Lévy et al., 2010; Arbic et al., 2013). The do-  
 512 main size of our model has allowed us to partially resolve submesoscale turbulence and  
 513 the inverse energy cascade associated with it. By coupling the channel model to a full  
 514 biogeochemical model, we have examined the relative importance of vertical eddy iron  
 515 transport associated with meso- and submeso-scale turbulence in the open Southern Ocean,  
 516 where understanding has primarily relied on vertical diffusion and mixing-layer entrain-  
 517 ment framework (Bowie et al., 2009; Tagliabue et al., 2012, 2014; Llort et al., 2015, 2019).

518 In order to quantify the temporal and spatial scales at which eddy transport was  
 519 dominant, we took the frequency-wavenumber cross spectra of  $w$  and Fe. The spectra  
 520 at depths below the wintertime maximum of MLD (Fig. 9) showed two things: i) most  
 521 of the vertical transport is at scales larger than the Rossby deformation radius and ii)  
 522 internal waves contribute to no net transport of iron. The first point already implies that  
 523 mesoscale turbulence is the dominant contributor to vertical iron transport below the  
 524 ML (Fig. 8). Considering that different dynamics can have similar spatial and tempo-

525 ral scales, we further dynamically decomposed the eddy transport into its balanced and  
 526 unbalanced component using the generalized Omega equation (Giordani & Planton, 2000)  
 527 based on the assumption that mesoscale turbulence is associated with Rossby numbers  
 528 smaller than unity (McWilliams, 2016; Lévy et al., 2018). At depths below the mixed  
 529 layer, we found that the balanced motion accounts for more than half of the total ver-  
 530 tical iron transport where as within the mixed layer where stratification is low and MLI  
 531 is active (Boccaletti et al., 2007), the relative contribution by unbalanced motion increases  
 532 significantly (Fig 10d).

533 Although our wintertime biogeochemical consumption of iron is within the bounds  
 534 of observations, it is too low during summer; Ellwood et al. (2008); Bowie et al. (2009)  
 535 estimate it to be on the order of  $100 \mu\text{mol m}^2 \text{yr}^{-1}$  while as it is roughly  $35 \mu\text{mol m}^2 \text{yr}^{-1}$   
 536 integrating over the top 100 m in our model (Fig. 8b; green dotted line). Due to the lack  
 537 of pelagic community transition, our bloom is too sharp and is insufficiently sustained  
 538 over the summer (Appendix B, Fig. B1). As the ecosystem is iron limited year round  
 539 (J. K. Moore et al., 2013), one approach for increasing summer productivity in future  
 540 work may be to reduce iron requirements for small phytoplankton, viz. have separate  
 541 iron-to-phosphate ratio per species (Table A2). Although Dutkiewicz et al. (2009), con-  
 542 figured to represent the global ecosystem, did not allow for this, it is likely that phyto-  
 543 plankton adaptation has occurred in the Southern Ocean where small phytoplankton should  
 544 have advantage under low iron concentrations (Tagliabue et al., 2014). Our sharp bloom  
 545 should not affect our results qualitatively, however, as an increase in summertime bio-  
 546 logical consumption would result in larger vertical gradients of iron; assuming that mesoscale  
 547 stirring is related to the background gradient of iron ( $w'Fe' \sim \kappa \frac{dFe}{dz}$  where  $\kappa$  here is the  
 548 eddy diffusivity), mesoscale eddy iron transport would only increase.

549 Our findings, which emphasize the importance of eddy iron transport, are compli-  
 550 mentary to Freilich and Mahadevan (2019, Figs. 6, 7e in their paper) in which they show  
 551 that isopycnal (sub)mesoscale stirring of nutrients increases with resolution. Rosso et  
 552 al. (2014, 2016), using an idealized biogeochemical model based on an exponential de-  
 553 cay rate of iron, also argued for the importance of submesoscale iron transport in the  
 554 Kerguelen Plateau region. Enhanced submesoscale turbulence in their model occurred  
 555 due to flow-bathymetric interaction, resulting in a hot-spot of vertical iron transport down-  
 556 stream of the Kerguelen Islands. Away from bathymetric features, however, they showed  
 557 that submesoscale eddy transport of iron was weak. Balwada et al. (2018) using a sub-

558 mesoscale permitting zonal re-entrant model with a topographic ridge showed that down-  
559 stream of the ridge, vertical tracer transport was enhanced at higher frequency and wavenum-  
560 ber. Both studies imply that surface submesoscale turbulence is enhanced due to flow-  
561 bathymetric interaction. Since our model has flat bottom, we do not have geographical  
562 hot spots of submesoscale turbulence. We argue that in the zonal-mean sense away from  
563 bathymetric features in the open Southern Ocean, it is the *local* effect near the surface  
564 where MLI is active and *remote* effect, i.e. mesoscale eddies energized through inverse  
565 energy cascade, at depths that dominate the eddy iron transport to the surface. In other  
566 words, our results suggest that it is first-order importance to get the energetics of the  
567 mesoscale field right either through resolving or parametrizing the inverse energy cas-  
568 cade due to submesoscale baroclinic instabilities for estimating iron and tracer transport.  
569 Current generation of eddy parametrizations do not incorporate the effect of inverse en-  
570 ergy cascade; however, parameterizing this energy transfer is an active area of research  
571 (e.g. Kitsios et al., 2013; Jansen et al., 2015; Anstey & Zanna, 2017; Bolton & Zanna,  
572 2019; Bachman, 2019). Due to computational constraints, GCMs at mesoscale permit-  
573 ting resolutions will continue to be invaluable tools to investigate the coupling between  
574 physics and biogeochemistry on a global scale.

575       Based on modelling studies with higher spatial resolution (Molemaker et al., 2010;  
576 Smith et al., 2016; Brannigan et al., 2017; Balwada et al., 2018), it is obvious that even  
577 at 2 km resolution, which is state of the art coupled to a full biogeochemical model, our  
578 turbulent field is not numerically converged; we would expect both the meso- and submeso-  
579 scale tracer transport to further increase with resolution. It would be interesting to see  
580 how the forward cascade of KE due to the transition from geostrophic to three-dimensional  
581 turbulence if otherwise resolved on the scales of  $O(100-1\text{ m})$ , and its non-linear interac-  
582 tion with the mean flow (Molemaker et al., 2005; Levy & Martin, 2013), would impact  
583 eddy iron transport in even high-resolution simulations.

584

## Appendix A Model boundary conditions

585

### A1 Physical forcing

The monthly structure of wind stress profile (Fig. A1) inspired by Sinha and Abernathey (2016) takes the mathematical form,

$$\tau(t, y) = \begin{cases} \left(0.15 - 0.05 \sin \left[\frac{\pi}{3}(t - 0.5)\right]\right) \sin^2 \left[\pi \frac{y - a/2}{L_y - a}\right] & , a/2 < y < L_y - a/2 \\ 0 & , y \leq a/2 \text{ or } y \geq L_y - a/2 \end{cases} \quad (\text{A1})$$

where  $L_y$  is the meridional extent of the domain,  $t \in [1, 2, \dots, 12]$  for each month and  $a = 100$  km. The westerly jet takes its maximum in the meridional center of the domain and tails off to zero towards the boundaries. The SST is relaxed to the profile (Fig. A1),

$$\theta^*(t, y) = \begin{cases} \frac{\theta_0 + \Delta\theta \sin \left[\frac{\pi}{6}(t - 1)\right]}{0.3L} y & , 0 < y < 0.3L_y \\ \frac{1.5}{0.4L} (y - 0.3L) + \theta_0 + \Delta\theta \sin \left[\frac{\pi}{6}(t - 1)\right] & , 0.3L_y < y < 0.7L_y \\ \frac{\theta_2 - \theta_1 + \Delta\theta \sin \left[\frac{\pi}{6}(t - 1)\right]}{0.3L} (y - L) + \theta_2 & , 0.7L_y < y < L_y \end{cases} \quad (\text{A2})$$

586

where  $(\theta_0, \theta_1, \theta_2) = (3.25, 4.75, 8)^\circ\text{C}$  and  $\Delta\theta = 1.75^\circ\text{C}$ . Other physical parameters

587

are listed in Table A1.

588

### A2 Biogeochemical forcing

The light, temperature and nutrient limitation to the phytoplankton growth rate are implemented as

$$\mu_i = \mu_{\max_i} \gamma_i^I \gamma_i^T \gamma_i^N \quad (\text{A1})$$

where  $\mu_{\max_i}$ ,  $\gamma_i^I$ ,  $\gamma_i^T$  and  $\gamma_i^N$  are the maximum growth rate of phytoplankton  $i$  (diatom:  $i = 1$ , small phytoplankton:  $i = 2$ ) and limitation factors by temperature, light and nutrients respectively. The full equation for the time evolution of phytoplankton is given in Dutkiewicz et al. (2009, eqns. A1-A5 in their paper) and parameter values in Table A2. Each limitation factor takes values between unity and zero ( $\gamma \in [0, 1]$ ) with one meaning optimum conditions for growth. Light limitation is calculated as

$$\gamma_i^I = \begin{cases} \min \left[ F_0^{-1} (1 - e^{-\kappa_{\text{par}} \text{PAR}}) e^{-\kappa_{\text{inh}} \text{PAR}}, 1 \right] & , \text{PAR} > 1 \\ 0 & , \text{PAR} < 1 \end{cases} \quad (\text{A2})$$

where PAR is the photosynthetically available radiation in units of  $[\mu\text{Ein m}^{-2} \text{s}^{-1}]$  and

$$F_0 = \frac{\kappa_{\text{par}}}{\kappa_{\text{par}} + \kappa_{\text{inh}}} e^{\frac{\kappa_{\text{inh}}}{\kappa_{\text{par}}} \log \left[ \frac{\kappa_{\text{inh}}}{\kappa_{\text{par}} + \kappa_{\text{inh}}} \right]}.$$

Parameter	Value	Units
Horizontal resolution	20, 5, 2	km
Time step	1800, 150, 40	s
Spin up	200, 35 (from 20 km), 4.5 (from 5 km)	years
SST relaxation piston velocity	1/3	m day <sup>-1</sup>
Redi diffusivity	200	m <sup>2</sup> s <sup>-1</sup>
*Typical ML frontal width ( $L_f$ )	2	km
*MLI efficiency ( $C_e$ )	0.07	
*MLI time scale ( $\tau$ )	$2 \times 10^{-6}$ (5.8)	s <sup>-1</sup> (day)
*Maximum grid-scale ( $L_{\max}$ )	110	km

**Table A1.** Physical parameter values for each resolution. The piston velocity is defined as the top grid cell height divided by the relaxation time scale. The parameters with (\*) are only used for the 20 km run with MLI parametrization turned on. Further information on other variables will be left to Balwada et al. (2018, Supplementary Material Table 1).

Figure A2d,e shows behaviour of  $\gamma^I$  and  $\mu_{\max}\gamma^I$  in the range of PAR used in our model; the light limitation factors ( $\gamma^I$ ) are in the range of values presented by Dutkiewicz et al. (2009, Fig. 1e in their paper. Note the difference in the units; 10 Ein m<sup>-2</sup> d<sup>-1</sup> corresponds roughly to 116  $\mu$ Ein m<sup>-2</sup> s<sup>-1</sup>). The temperature limitation is kept the same for both phytoplankton species and takes a similar formulation to the Arrhenius equation (e.g. Geider, 1987; Brown et al., 2004; Kremer et al., 2017)

$$\gamma_i^T = T_0 \max\left[e^{-T_{\text{Ae}}\left(\frac{1}{T+T_{\text{kel}}} - \frac{1}{T_{\text{ref}}}\right)}, 10^{-10}\right] \quad (\text{A3})$$

where  $T$  is the local temperature in Celsius and the reference temperature  $T_{\text{ref}}$  defines the temperature at which  $\gamma^T = T_0$ . Lastly, the nutrient limitation factor is defined by the most limiting nutrient

$$\gamma_i^N = \min(N_{ij}^{\text{lim}}) \quad (\text{A4})$$

589 where  $N_{ij}^{\text{lim}} = \frac{N_j}{N_j + \kappa_{N_{ij}}}$  with  $N_j$  and  $\kappa_{N_{ij}}$  being the concentration of nutrient  $j$  and half-  
 590 saturation constant of nutrient  $j$  for phytoplankton  $i$  respectively. Figure A3 shows that  
 591  $N_{i\text{Fe}}^{\text{lim}}$  is always smallest amongst the nutrient limitation factors per phytoplankton, viz.  
 592  $\gamma_i^N = N_{i\text{Fe}}^{\text{lim}} = \frac{\text{Fe}}{\text{Fe} + \kappa_{i\text{Fe}}}$ , with a large dip during Oct.-Dec. when the climax of spring  
 593 bloom takes place. We also see the effect of diatoms having larger half saturation con-  
 594 stants;  $N^{\text{lim}}$  is lower for diatoms than small phytoplankton for each nutrient year round.  
 595 Further details of the variables and notations will be left to Dutkiewicz et al. (2009), which  
 596 the reader should refer to.

## 597 **Appendix B Seasonality in phytoplankton biomass**

598 The resulting seasonal cycle of vertically integrated phytoplankton biomass ( $\langle C_p \rangle$ )  
 599 is shown in Fig. B1 along with estimates from biogeochemical (BGC) Argo floats deployed  
 600 by the SOCCOM and SOCLIM projects (Johnson et al., 2017; Riser et al., 2018; Ley-  
 601 marie et al., 2018), and surface biomass concentration ( $C_p$ ) from the carbon-based pro-  
 602 duction model (CbPM; Westberry et al., 2008). Details of the BGC-Argo float dataset  
 603 and quality control will be left to Uchida et al. (submitted). The apex of the spring bloom  
 604 ( $\langle C_p \rangle$  maximum) in our 2 km run occurs roughly 1-2 months earlier than the Argo es-  
 605 timates (Fig. B1a,c,e), which is likely due to the over dominance of diatoms in our ecosys-  
 606 tem. The basin-wide maximum of  $\langle C_p \rangle$ , however, aligns meridionally with where our ACC  
 607 lies, indicated by the maximum in KE (Fig. B1b). This is consistent with the Argo es-  
 608 timate between 45S-60S (Fig. B1c,d). The maximum in KE around 40S is due to the  
 609 Agulhas Current and downstream of the Drake Passage. Although we only analyze the  
 610 meridional extent of  $y \in [600, 1400]$  km in our study, it is also interesting to note the  
 611 indentation in the southward progression of apex around  $y \sim 300$  km (Fig. B1a), which  
 612 is likely due to our cosine-squared form of wind stress (eqn. A1); the ML shoals towards  
 613 the north and south boundaries (Fig. B1b). The ML at the very south is deep due to  
 614 year-round convection at the boundary wall.

## 615 **Appendix C Omega equation**

The right hand side of eqn. 3.2.1 is  $\mathbf{Q} = \mathbf{Q}_{\text{tw}} + \mathbf{Q}_{\text{da}}$  where

$$\mathbf{Q}_{\text{tw}} = -2 \left( \frac{\partial \mathbf{u}}{\partial x} \cdot \nabla b, \frac{\partial \mathbf{u}}{\partial y} \cdot \nabla b \right) \quad (\text{C1})$$

$$\mathbf{Q}_{\text{da}} = f \left( \frac{\partial v}{\partial x} \frac{\partial u_a}{\partial z} - \frac{\partial u}{\partial x} \frac{\partial v_a}{\partial z}, \frac{\partial v}{\partial y} \frac{\partial u_a}{\partial z} - \frac{\partial u}{\partial y} \frac{\partial v_a}{\partial z} \right) \quad (\text{C2})$$



Parameter	Symbol	Value	Units
Phytoplankton max. growth rates	$\mu_{\max(1,2)}$	1/1.24, 1/1.8	days <sup>-1</sup>
Mortality rates	$m_{P(1,2)}$	15, 12	days
	$m_{Z(1,2)}$	40, 40	days
Zooplankton max. grazing rates	$g_{\max_a}$	2.8	days
	$g_{\max_b}$	16.8	days
Half-saturation constants	$\kappa_{\text{PO}_4(1,2)}$	0.035, 0.015	mmole m <sup>-3</sup>
	$\kappa_{\text{NO}_3(i)}$	$\kappa_{\text{PO}_4(i)} \times r_{\text{N:P}}$	mmole m <sup>-3</sup>
	$\kappa_{\text{Fe}_T(i)}$	$\kappa_{\text{PO}_4(i)} \times r_{\text{Fe:P}}$	mmole m <sup>-3</sup>
	$\kappa_{\text{Si}(1)}$	$\kappa_{\text{PO}_4(1)} \times r_{\text{Si:P}}$	mmole m <sup>-3</sup>
Phytoplankton elemental ratios	$r_{\text{N:P}}$	16	
	$r_{\text{Si:P}}$	16	
	$r_{\text{Fe:P}}$	10 <sup>-3</sup>	
PAR saturation constants	$\kappa_{\text{par}(1,2)}$	0.018, 0.01	( $\mu\text{Ein m}^{-2} \text{s}^{-1}$ ) <sup>-1</sup>
PAR inhibition constants	$\kappa_{\text{inh}(1,2)}$	$1.05 \times 10^{-3}$ , $5.9 \times 10^{-3}$	( $\mu\text{Ein m}^{-2} \text{s}^{-1}$ ) <sup>-1</sup>
Normalization constant	$T_0$	0.589	
Activation temperature	$T_{\text{Ae}}$	4040	K
Absolute zero temperature	$T_{\text{kel}}$	273.15	K
Reference temperature	$T_{\text{ref}}$	277.15	K
Nutrition relaxation time scale		30	days

**Table A2.** The two-species Darwin parameter values used in our configuration where  $i(= 1, 2)$  correspond to Diatoms which have an additional silicate dependence and small phytoplankton respectively. In Dutkiewicz et al. (2009), the units of the half-saturation constants are in [ $\mu\text{M}$ ] (micromole per liter), equivalent to [mmole m<sup>-3</sup>].

616 Assuming the total flow to be in geostrophic balance ( $\mathbf{u} = \mathbf{u}_g = \frac{\hat{z}}{f} \times \nabla_h \Phi$ ) reduces  
 617 eqn. (3.2.1) to the quasi-geostrophic Omega equation (Hoskins et al., 1978). The ageostrophic  
 618 velocities were defined as the difference between the total and geostrophic velocity, i.e.  $\mathbf{u}_a =$   
 619  $\mathbf{u} - \mathbf{u}_g$ . We have neglected the terms  $\mathbf{Q}_{th}$  and  $\mathbf{Q}_{dm}$  in Giordani and Planton (2000), which  
 620 correspond to dissipation due to Leith viscosity (Bachman et al., 2017), KPP (Large et  
 621 al., 1994) and/or bottom drag in our case as we use snapshot outputs to invert eqn. (3.2.1).  
 622 We find the Omega equation performs well enough for our purpose of decomposing the  
 623 flow without incorporating these terms (Fig. 6).

We solve for eqn. (3.2.1) in the horizontal wavenumber Fourier space at each vertical level and time, i.e.

$$-\kappa^2 N^2 \hat{w} + f_0^2 \frac{\partial^2 \hat{w}}{\partial z^2} = ik\beta \hat{b} + (ik\hat{Q}^x + il\hat{Q}^y) \quad (C3)$$

624 where  $\kappa = (k, l)$  is the horizontal wavenumber vector using the `xomega` Python pack-  
 625 age we developed (<https://xomega.readthedocs.io/en/latest/>).

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 630 simulation outputs for 15-daily snapshot and monthly averaged outputs of physical vari-  
 631 ables ( $\mathbf{v}, \theta, \Phi$ ) are available on Pangeo ([https://pangeo-data.github.io/pangeo-datastore/](https://pangeo-data.github.io/pangeo-datastore/master/ocean/channel.html)  
 632 [master/ocean/channel.html](https://pangeo-data.github.io/pangeo-datastore/master/ocean/channel.html)). For other variables, please contact the leading author.  
 633 The World Ocean Atlas data was acquired via [https://data.nodc.noaa.gov/thredds/](https://data.nodc.noaa.gov/thredds/dodsC/woa/WOA13/DATA/)  
 634 [dodsC/woa/WOA13/DATA/](https://data.nodc.noaa.gov/thredds/dodsC/woa/WOA13/DATA/), BSOSE via [http://sose.ucsd.edu/bsose\\_solution\\_Iter105](http://sose.ucsd.edu/bsose_solution_Iter105.html)  
 635 [.html](http://sose.ucsd.edu/bsose_solution_Iter105.html), and SeaWiFS PAR data via [https://oceandata.sci.gsfc.nasa.gov/SeaWiFS/](https://oceandata.sci.gsfc.nasa.gov/SeaWiFS/Mapped/Daily/9km/par/)  
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 644 [.oregonstate.edu/1080.by.2160.8day.hdf.carbon2.m.php](http://orca.science.oregonstate.edu/1080.by.2160.8day.hdf.carbon2.m.php)).

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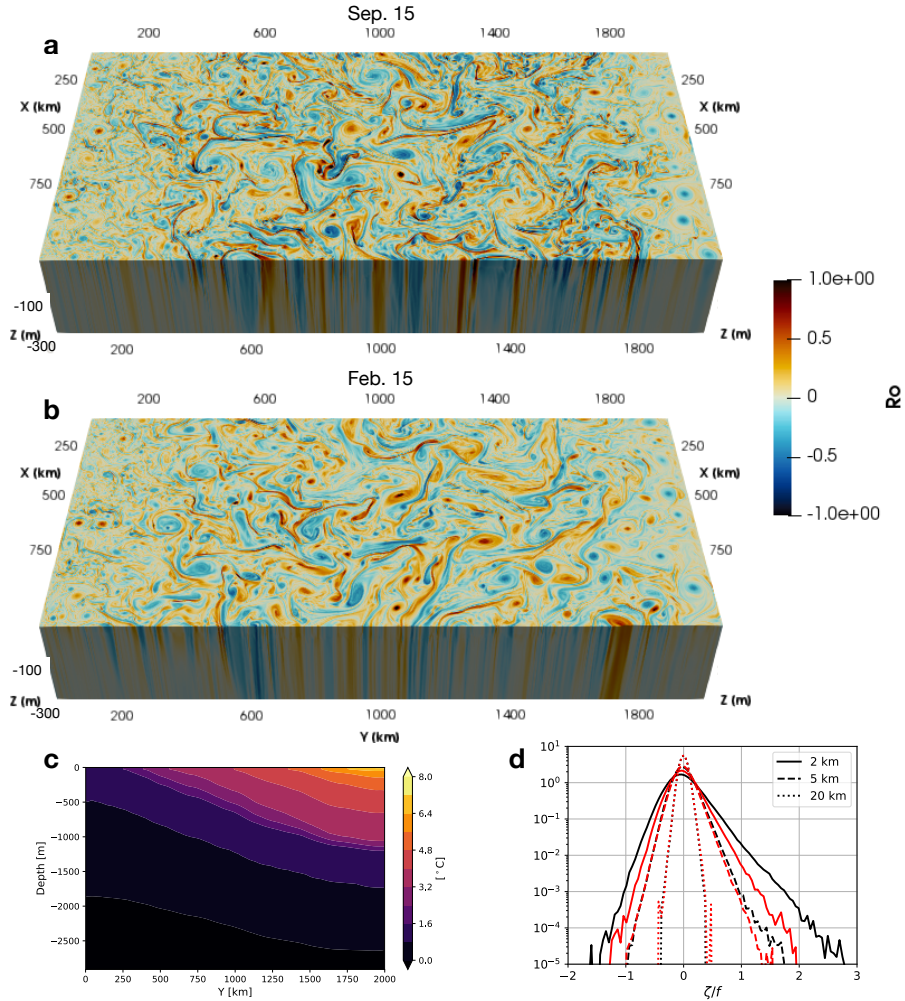
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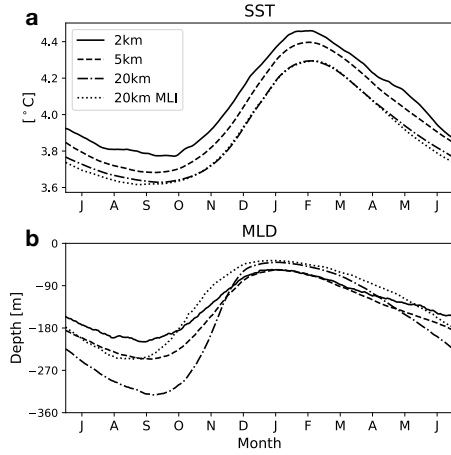
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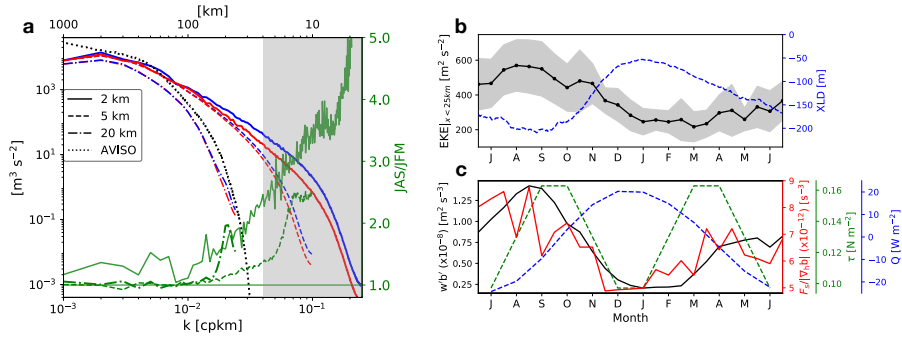
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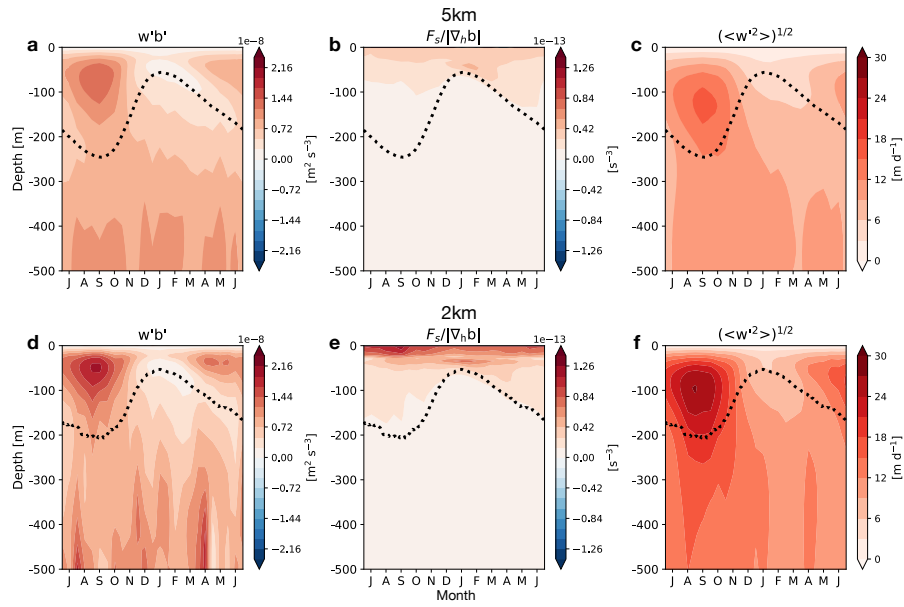
**Figure 1.** Relative vorticity normalized by the local Coriolis parameter ( $Ro = \frac{\zeta}{f}$ ) for the 2 km run on September 15 **a** and February 15 **b** in the top 300 m. **c** The zonal-annual mean stratification from the 2 km run plotted against depth and meridional distance. **d** Seasonal probability density function of  $Ro$  for each resolution is shown in. Winter (July, August, September) is shown in black and summer (January, February, March) in red.



**Figure 2.** The spatial average over the meridional extent of  $y \in [600, 1400]$  km of daily-averaged SST from the 2 km run **a**. Note that the month axis starts from July, austral winter. **b** The spatial mean of daily-averaged MLD for each run.

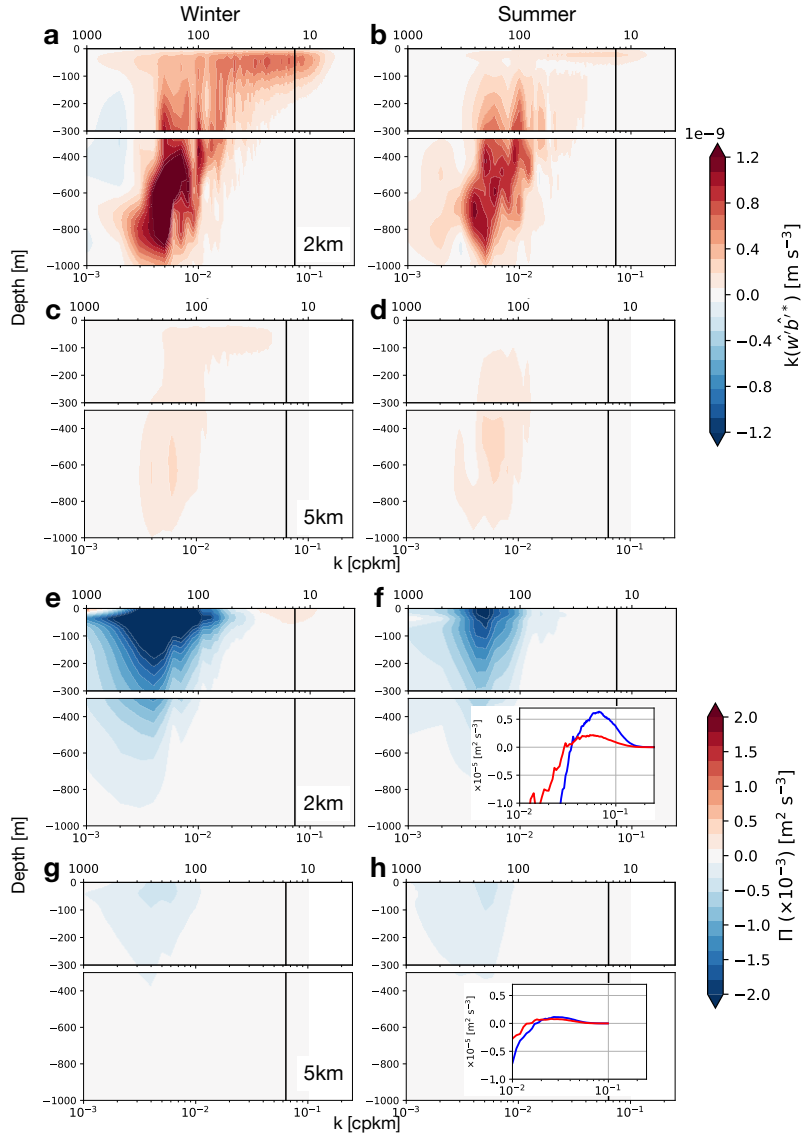


**Figure 3.** Seasonal and meridional mean of the zonal wavenumber KE power spectra for the 2 km (solid), 5 km (dashed) and non-parametrized 20 km run (dotted-dash), and climatology of AVISO (dotted) **a**. The  $x$  axis at the top shows the wavelength each wavenumber corresponds to. Red (blue) indicates winter (summer). The green lines show the ratio of austral winter over summer for the 2 km (solid), 5 km (dashed), and 20 km run (dotted-dash) and the grey shading shows scales below 25 km. **b** The time series of surface KE at scales smaller than 25 km and MLD (blue dashed) depth averaged over the top 100 m from the 2 km run. **c** Baroclinic energy conversion rate (black solid) and frontogenesis function (red solid) depth averaged over the top 100 m, surface wind stress (green dashed) and heat flux (orange dashed).

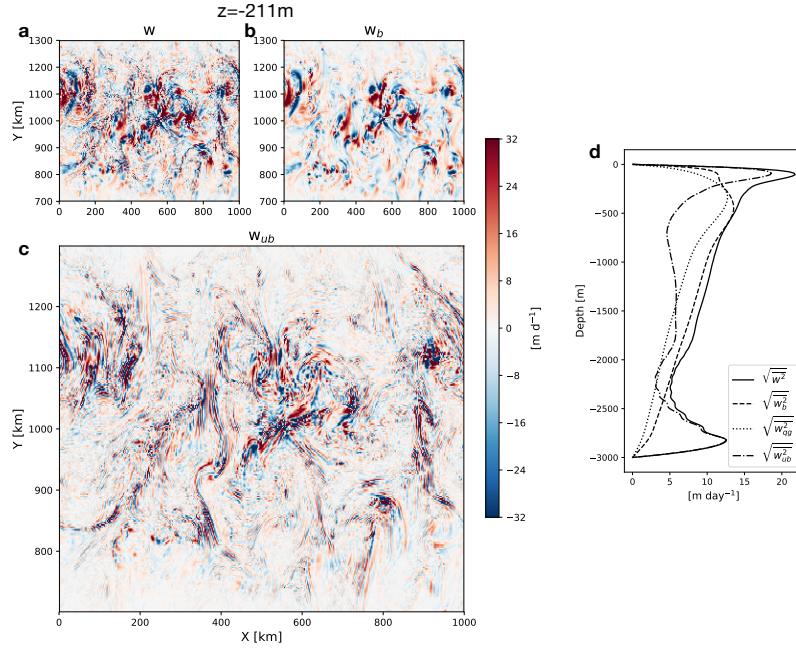


**Figure 4.** Hovmöller diagram of baroclinic energy conversion rate ( $w'b'$ ) averaged spatially over  $y \in [600, 1400]$  km (a,d). b,e Monthly climatology of the frontogenesis function normalized by the horizontal buoyancy gradient in  $\text{s}^{-3}$ , and root-mean square of  $w$  [ $\text{m d}^{-1}$ ] (c,f). Each variable is shown for the two resolution 5 and 2 km runs and MLD in black dotted lines.

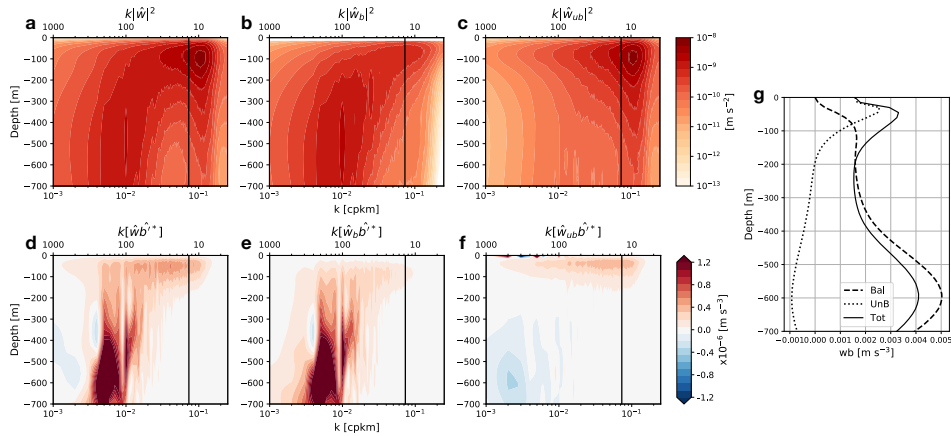




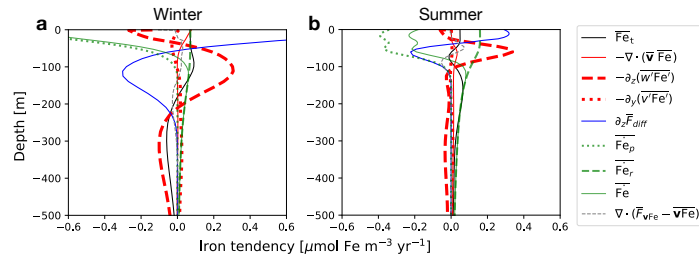
**Figure 5.** Seasonal-meridional mean over  $y \in [600, 1400]$  km of the zonal-wavenumber cross spectra of vertical velocity and buoyancy ( $\mathcal{R}[\hat{w}'\hat{b}'^*(k)]$ ) in variance preserving form for winter **a** and summer **b** for the 2- and 5-km (**c,d**) run. The spatial scale corresponding to each wavenumber is shown on the top  $x$  axes of each panel in [km]. **e-h** The spectral flux of total KE ( $\Pi(k)$ ) is shown in the same manner. The black solid lines show the Rossby deformation wavenumber. The lower-right panel is a zoomed-in figure of the KE cascade at scales below 100 km averaged over the top 100 m with winter (summer) shown in blue (red).



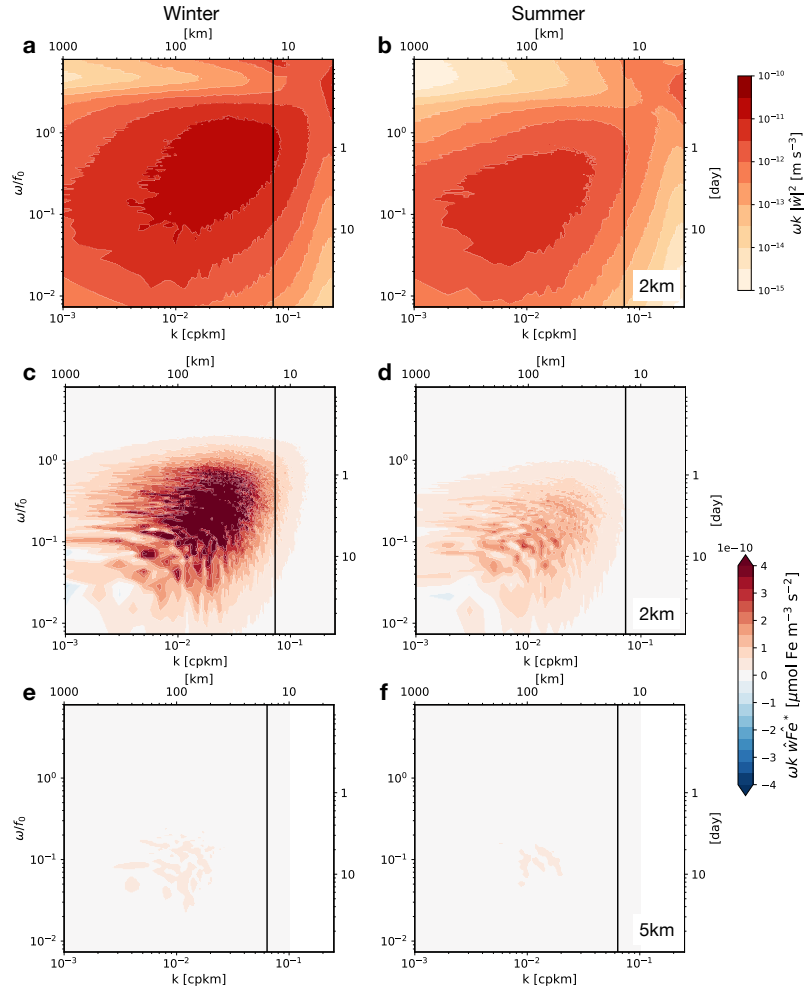
**Figure 6.** Snapshot of the total vertical velocity ( $w$ ; **a**) on August 15, balanced ( $w_b$ ; **b**) and unbalanced ( $w_{ub}$ ; **c**) at the depth of  $z = -211$  m. **d** The vertical profile of the root-mean square of each term along with an inversion using quasi-geostrophic approximation ( $w_{qg}$ ). The overbar ( $\overline{\cdot}$ ) indicates the spatial averaging.



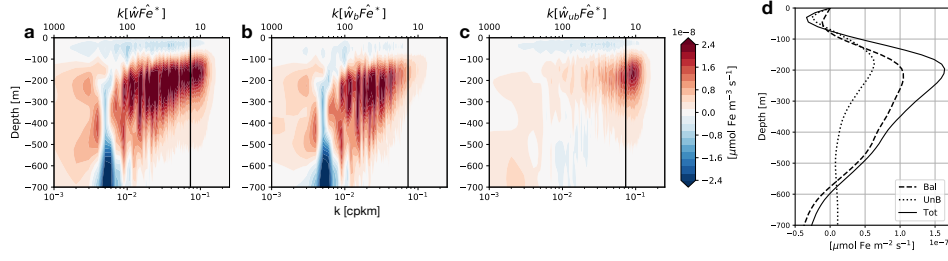
**Figure 7.** The zonal wavenumber power spectra of total vertical velocity ( $w$ ; **a**), balanced ( $w_b$ ; **b**) and unbalanced ( $w_{ub}$ ; **c**) in variance preserving form. The spatial scales corresponding to the wavenumbers are again shown on the top  $x$  axes in [km]. **d-f** Zonal wavenumber cross spectra of vertical velocity and buoyancy ( $\mathcal{R}[\hat{w}\hat{b}^{*}]$ ) in the same manner. The black solid lines show the Rossby deformation wavenumber. **g** Profile of vertical buoyancy flux associated with each component.



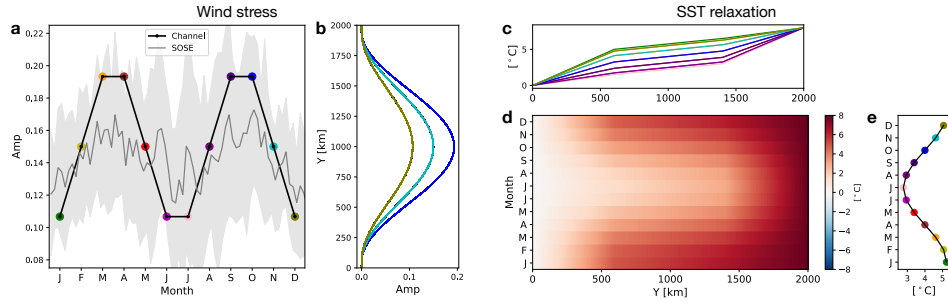
**Figure 8.** The iron budget for winter (July, August, September; **a**) and summer (January, February, March; **b**) over the meridional extent of  $y \in [600, 1400]$  km from the 2 km run. The residual term in grey dashed lines comes from the tracer advective flux scheme which prevents tracer concentrations from taking negative values.



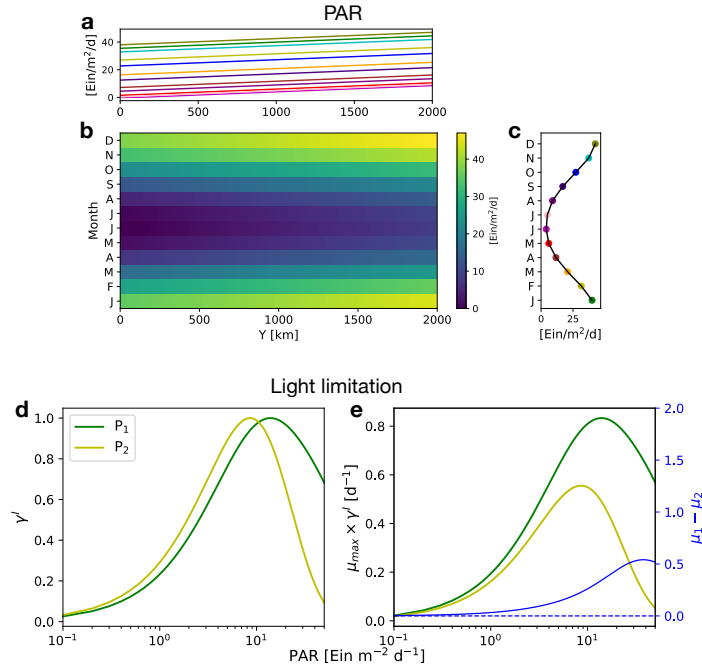
**Figure 9.** Frequency-zonal wavenumber power spectra of  $w$  (a,b) and cross spectra of  $w$  and Fe (c,d) in variance preserving form from the 2-km run at  $z = -211$  m depth using hourly snapshot outputs for winter (July, August, September; left) and summer (January, February, March; right). Cross spectra for the 5-km run are shown in e,f. Frequency is normalized by the Coriolis parameter at the center of the domain. The top and right  $x$ - $y$  axes in each panel show the corresponding spatial scale in [km] and dimensionalized frequency in the units of [day]. The spectra were averaged over the meridional extent  $y \in [600, 1400]$  km. The black solid lines show the Rossby deformation wavenumber.



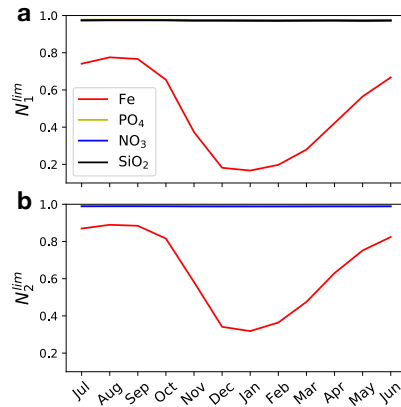
**Figure 10.** The variance preserving zonal-wavenumber cross spectra of vertical velocity and iron averaged over the meridional extent of  $y \in [800, 1200]$  km and winter (JAS) for total **a**, balanced **b**, and unbalanced vertical velocity **c**. The black solid lines show the Rossby deformation wavenumber. **d** The vertical profile of iron transport attributed to each component.



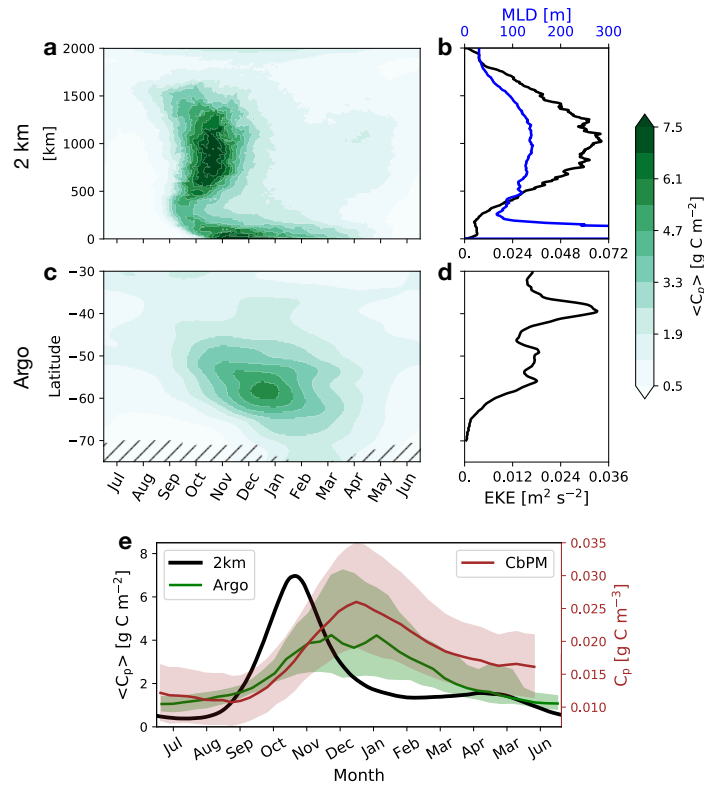
**Figure A1.** The monthly surface wind stress amplitude for each month in our model is plotted against the annual climatology of the Southern Ocean State Estimate (SOSE; verdy2017data wind stress averaged zonally between 50-56S **a**. The grey shadings are the standard deviation of SOSE. **b** The meridional profile of our westerly jet with the colors corresponding to them in panel **a**. **d** SST relaxation profiles are plotted against month and meridional extent and panel **c** is showing the same thing along temperature and meridional extent. **e** The meridional average of **c** with colors corresponding with each other.



**Figure A2.** Surface PAR in our runs plotted against month and meridional extent **b**. **c** The meridional mean of **a** and **b** with the colors of **a** and **c** correspond with each other. **d** The light limitation factor for each plankton (eqn. A2; Diatoms ( $P_1$ ) in green and small phytoplankton ( $P_2$ ) in yellow) and it multiplied by the maximum growth rate in **e**. The blue line shows the difference between the two ( $\mu_1 - \mu_2$ ).



**Figure A3.** Nutrient limitation factor (eqn. A4) for diatoms ( $i = 1$ ; **a**) and small phytoplankton ( $i = 2$ ; **b**) using monthly-mean nutrient outputs from the 2 km run averaged over the top 100 m.



**Figure B1.** Hovmöller diagram of  $\langle C_p \rangle$  for the 2 km run **a**, and from BGC Argo floats **c**. The hatch indicates grids where there were less than 10 Argo profiles over the five years of deployment by SOCCOM and SOCLIM. **b,d** Zonal mean of surface KE and MLD from the 2 km run and AVISO plotted against meridional axes. **e** Spatial and daily median of  $\langle C_p \rangle$  for the 2 km run between  $y \in [600, 1400]$  km (black), 10-daily median BGC Argo (green), and eight-day averaged  $C_p$  outputs from CbPM (brown) between 45S-60S. Note the difference in the units. As CbPM relies on satellite Chlorophyll observations, the month of June lacks data due to poor light conditions. The green and brown shading indicated the zonal and daily interquartile range.