

Air-Sea CO₂ Fluxes Localized By Topography in a Southern Ocean Channel

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Key Points:

- We examine the localized patterns of air-sea CO₂ fluxes in an idealized Southern Ocean-like model with simple biogeochemistry.
- We find intense sea-air CO₂ fluxes upstream of seafloor topography driven by anomalous advection of inorganic carbon.
- Due to the topography, uncertainty in the flux is highly sensitive to sampling network design.

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Abstract

Air-sea exchange of carbon dioxide (CO_2) in the Southern Ocean plays an important role in the global carbon budget. Previous studies have suggested that flow around topographic features of the Southern Ocean enhances the upward supply of carbon from the deep to the surface, influencing air-sea CO_2 exchange. Here, we investigate the role of seafloor topography on the transport of carbon and associated air-sea CO_2 flux in an idealized channel model. We find elevated CO_2 outgassing downstream of a seafloor ridge, driven by anomalous advection of dissolved inorganic carbon. Argo-like Lagrangian particles in our channel model sample heterogeneously in the vicinity of the seafloor ridge, which could impact float-based estimates of CO_2 flux.

Plain Language Summary

The Southern Ocean, the ocean surrounding Antarctica, contributes significantly to carbon exchange between the global ocean and the atmosphere, which in turn matters for climate change. Here, we use a simplified model of the Southern Ocean to see how mountain ranges on the sea floor influence the carbon exchange at the ocean-atmosphere interface. We find that the seafloor mountain ranges lead to more carbon exchange. Floating carbon sensors in our model ocean may under or over sample the water near the mountains and this can affect the carbon exchange that they report.

1 Introduction

The Southern Ocean is an active driver in the global cycling of carbon dioxide (CO_2). Studies based on coarse-resolution ocean general circulation models suggest that the Southern Ocean carbon cycle is characterized by the surfacing of old, respired carbon from depth at high latitudes and the subduction of anthropogenic carbon driven by the meridional overturning circulation from the surface into the interior at mid latitudes (Mikaloff Fletcher et al., 2006, 2007). However, observations of the resulting air-sea CO_2 fluxes from these physical circulation processes are sparse in both space and time (Bakker et al., 2016), and this has limited our ability to accurately quantify the Southern Ocean's role in the global carbon budget. New observations from autonomous floats equipped with pH sensors as part of the Southern Ocean Carbon and Climate Observations and Modeling (SOC-COM) program suggest that the outgassing of respired carbon in high latitudes has pre-

45 viously been underestimated (Gray et al., 2018; Bushinsky et al., 2019), suggesting there
46 is more work to be done to constrain the air-sea carbon fluxes.

47 One contributing factor to the uncertainty in the Southern Ocean carbon budget
48 is spatial variability in the air-sea CO₂ flux that is engendered by regional variations in
49 the physical circulation. While the canonical view of Southern Ocean circulation is an
50 annular circumpolar current with a broad region of surface divergence and upwelling at
51 $\sim 55^\circ\text{S}$ and convergence and subduction at $\sim 40^\circ\text{S}$ (Speer et al., 2000), current literature
52 highlights the non-annular nature of the circumpolar current (Rintoul, 2018) and asso-
53 ciated overturning circulation (Youngs & Flierl, 2023). Seafloor topographic features such
54 as ridges create standing meanders in the current and drive localized upwelling (e.g., Tam-
55 sitt et al., 2017; Youngs & Flierl, 2023), and it is thought that these topographic features
56 may play an important role in carbon fluxes. High resolution ocean circulation and bio-
57 geochemical modeling studies suggest that standing meanders contribute to southward
58 transport of anthropogenic carbon (Ito et al., 2010), and that intensified residual upwelling
59 downstream of regional topographic features provides an important conduit for deep, nat-
60 ural carbon to enter the Southern Ocean surface (Brady et al., 2021). Despite the po-
61 tentially important role that these regional topographic features play in the global car-
62 bon budget, no study has directly quantified the influence of seafloor topography on South-
63 ern Ocean air-sea CO₂ flux nor addressed the potential effects these features may have
64 on Lagrangian observations of the Southern Ocean.

65 Here, we use an idealized, high-resolution ocean general circulation and biogeochem-
66 ical model to assess the role of seafloor topography in Southern Ocean air-sea CO₂ fluxes
67 and the ability to quantify these fluxes via Lagrangian observations. Our study demon-
68 strates that seafloor topography has a substantial impact on local CO₂ flux via topography-
69 driven advection of dissolved inorganic carbon (DIC). Lagrangian particles tend to het-
70 erogeneously sample the surface pCO₂ in the vicinity of topography, and this can affect
71 estimates of average air-sea CO₂ fluxes over the region. In section 2, we present the meth-
72 ods used, in section 3 we present the results. In section 4 we discuss and conclude.

2 Methods

2.1 Model description

For this study, we use an idealized-geometry MITgcm ocean channel model (Youngs & Flierl, 2023) and couple it to a simple ocean biogeochemical model (Dutkiewicz et al., 2005; Lauderdale et al., 2016). The channel is 4000 km long and 2000 km wide with 10 km horizontal resolution (Figure 1) with a total depth of 4000 m with 32 vertical levels, from 10 m vertical grid spacing at the surface to 280 meters at the bottom. We represent seafloor topography using a 2000 m tall Gaussian ridge with a 200 km half-width, centered 800 km downstream of the channel entrance spanning the channel north to south (Figure 1). The domain is periodic with the outflow in the east reentering in the western boundary and free-slip walls at the north and the south. The model is integrated using a 600 second time step, an exponentially varying diffusivity ($0.01 \text{ m}^2 \text{ s}^{-1}$ to $1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$), and linear bottom drag with a drag coefficient of $1.1 \times 10^{-3} \text{ m s}^{-1}$. The wind stress is a cosine profile with a maximum value of 0.15 N m^{-2} at the center of the domain and zero wind stress at the sides (SI Fig. 1). The salinity is set at 35 PSU and not allowed to vary.

We employ the DIC package from MITgcm to represent biogeochemistry in our model (Dutkiewicz et al., 2005; Lauderdale et al., 2016). This model package carries alkalinity, DIC, dissolved organic phosphate, and phosphate as biogeochemical tracers, and represents biological uptake as a function of phosphate and light availability. Phosphate is fluxed vertically with remineralization and sinking (see more in the SI). The calcium carbonate formation is proportional to the organic phosphorous produced in the surface waters following the parameterization of Yamanaka and Tajika (1996), with sinking and dissolution (Dutkiewicz et al., 2005).

The rate of change of carbon in our model can be described by the following equation (Lauderdale et al., 2016)

$$\frac{\partial C_T}{\partial t} = \underbrace{-\nabla \cdot (\vec{u}C_T)}_{\text{Advection}} + \underbrace{\nabla \cdot (\kappa \nabla C_T)}_{\text{Diffusion}} - \underbrace{R_{C_T:P} S_{bio} - S_{CaCO_3}}_{\text{Biology}} - \underbrace{\frac{F_{CO_2}}{h}}_{\text{Air-sea fluxes}}, \quad (1)$$

where C_T is the concentration of total dissolved organic carbon, κ is the eddy diffusivity tensor, $R_{C_T:P}$ is the biological transformation between carbon and phosphorous and F_{CO_2} is the air-sea CO_2 flux, h is the mixed layer depth, S_{bio} represents the sources and

100 sinks of biogenic soft tissue, and S_{CaCO_3} represents the sources and sinks of biogenic car-
 101 bonate. Note that this equation neglects the dilution by freshwater fluxes, which in our
 102 case is appropriate due to a lack of salinity or freshwater forcing.

103 The model is initialized with a uniform surface ocean pCO_2 of 270 ppm with DIC
 104 and alkalinity at the northern boundary sponge region relaxed to prescribed DIC and
 105 alkalinity profiles based on GLODAPv2.2016 (Key et al., 2015; Lauvset et al., 2016) (SI
 106 F3), and spun up for 30 years for the biogeochemical and physical tracers to reach an
 107 approximate steady-state (table SI). At the end of the spin-up period, our model sim-
 108 ulates similar Southern Ocean-integrated pre-industrial air-sea CO_2 fluxes (0.1 mol m^{-2}
 109 yr^{-1}) as those estimated from more realistic model configurations ($0.13 \text{ mol m}^{-2} \text{ yr}^{-1}$)
 110 (e.g., Lovenduski et al., 2007).

111 2.2 Particle Tracking

112 We model idealized “Argo” float trajectories to estimate how well a biogeochem-
 113 ical Argo float array can sample the air-sea carbon fluxes as a function of float density.
 114 We use the Ocean Parcels package to track idealized Argo floats ([https://oceanparcels](https://oceanparcels.org/)
 115 [.org/](https://oceanparcels.org/)) (Lange & van Sebille, 2017). We release 800 floats spaced uniformly throughout
 116 the model domain. Real Argo floats park at 1000 m depth for 10 days between profiles,
 117 so in our simulations the particles are advected using daily-averaged velocities at 1000 m;
 118 they sample the surface ocean pCO_2 at their position every 10 days. Idealized floats are
 119 advected for 1 or 3 years. We take 100 random subsamples of each collection of ideal-
 120 ized floats with replacement. We run 4 collections of experiments: 10 floats for 1 year,
 121 33 floats for 1 year, 100 floats for 1 year, and 33 floats for 3 years. We use the randomly
 122 subsampled float data to create a climatology using objective mapping (e.g. Figure 3b).
 123 From the mapped pCO_2 , we calculate the air-sea carbon fluxes using the same equations
 124 used by the model (Wanninkhof, 1992).

125 Objective mapping is a commonly used and well-justified technique for mapping
 126 sparsely sampled data to estimate regional averages (Dong et al., 2008; Friedrich & Os-
 127 chlies, 2009; Reeve et al., 2016). We create climatologies of these samples using the or-
 128 dinary kriging method with the PyKriging Python package ([https://github.com/GeoStat-](https://github.com/GeoStat-Framework/PyKriging/)
 129 [Framework/PyKriging/](https://github.com/GeoStat-Framework/PyKriging/)). Here, the various terms for the Gaussian variogram are fit us-
 130 ing the data from the selected floats to create the most optimal map.

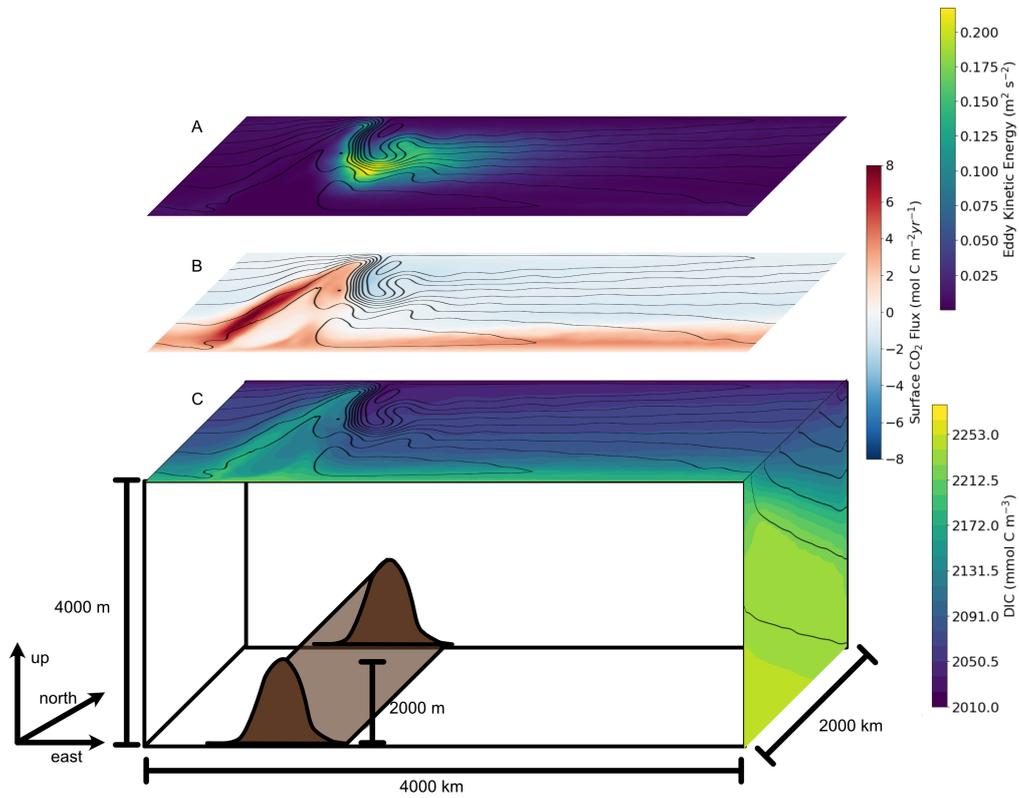


Figure 1. The model is a re-entrant channel forced with both a zonal wind and a relaxation to a meridional temperature gradient. Barotropic streamlines are shown with black contours on the top faces. Shading shows temporally-averaged (A) surface eddy kinetic energy, (B) surface carbon dioxide flux, and (C) dissolved inorganic carbon (DIC) concentration. In (C) the right edge shows temporally and zonally averaged DIC concentration with temperature (density) contoured in black. The model geometry is shown in C. A 2000 m tall undersea Gaussian ridge is centered at $x = 800$ km.

131 **3 Results**

132 **3.1 DIC budget**

133 We investigate the asymmetry of the carbon properties in the channel model. Both
 134 air-sea CO₂ flux and surface DIC concentration exhibit large zonal asymmetry, with en-
 135 hanced CO₂ outgassing and elevated surface DIC located just upstream of the under-
 136 sea ridge (Figure 1BC). Away from the influence of topography, our model exhibits mod-
 137 erate outgassing of CO₂ near the southern boundary, with weak uptake in the northern
 138 part of the domain (Figure 1B), which together contribute to an average flux of about
 139 $-0.07 \text{ mol C m}^{-2} \text{ yr}^{-1}$. At the latitudes of the topographic ridge, however, we find sea-
 140 air CO₂ fluxes that exceed $7 \text{ mol C m}^{-2} \text{ yr}^{-1}$ and outgassing that extends to the north-
 141 ern boundary of the domain, with an average flux of $0.8 \text{ mol C m}^{-2} \text{ yr}^{-1}$. The enhanced
 142 carbon flux is located in the region where the barotropic flow turns north as it approaches
 143 the ridge (Figure 1BC). This region is characterized by elevated surface DIC concentra-
 144 tions relative to the zonal mean for the domain (Figure 1C). We also show that as the
 145 wind stress forcing changes, the pCO₂ flux changes are driven by changes in advection
 146 of DIC not other terms like temperature forcing or changes in alkalinity (SI figure 5),
 147 highlighting the importance of the advection of DIC.

148 We investigate the drivers of the elevated surface ocean DIC upstream of the to-
 149 pographic ridge by quantifying the terms in Equation 1 averaged over the top 50 m. DIC
 150 advection tends to increase DIC upstream of the ridge, while sea-air CO₂ flux tends to
 151 decrease DIC in this same region (Figure 2A,B). In contrast, biological productivity tends
 152 to decrease DIC relatively uniformly over the domain, with only a slightly larger influ-
 153 ence upstream of the ridge, and DIC diffusion exhibits only a small influence on upper
 154 ocean DIC tendency across the domain (Figure 2C,D). The elevated net DIC advection
 155 upstream of the ridge is mostly driven by vertical advection (SI Figure 4), though the
 156 contribution from the horizontal advection of DIC is non-negligible, especially in the north-
 157 ern portion of the model domain (SI Figure 4). Thus, results from our DIC tendency bud-
 158 get suggest that enhanced vertical advection of DIC upstream of the ridge is responsi-
 159 ble for the locally elevated DIC, and by inference, the enhanced outgassing of CO₂ in
 160 this region. Our model also simulates elevated sea-air CO₂ flux and surface ocean DIC
 161 in the northern portion of model domain over the ridge, albeit with lower magnitudes
 162 than in the region upstream of the ridge (Figure 1). Here, the elevated DIC is driven by

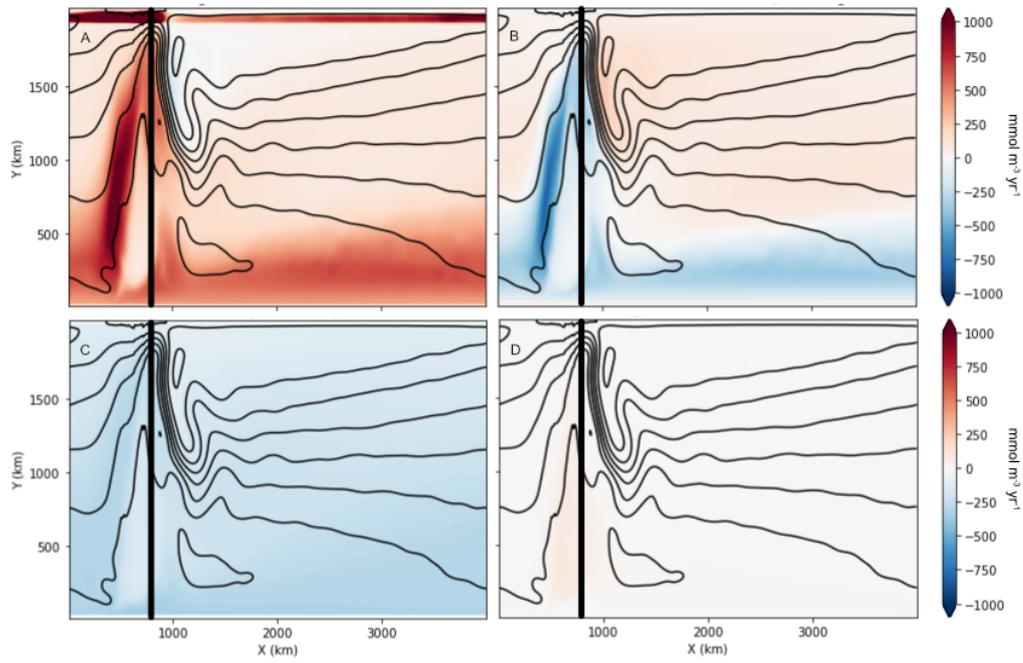


Figure 2. The drivers of the rate of change of DIC ($\frac{\partial C_T}{\partial t}$; $\text{mmol m}^{-3} \text{ yr}^{-1}$), as in Equation 1, averaged over the 20 year simulation and the top 50 m: (A) DIC advection, (B) sea-air flux of CO_2 , (C) biology, and (D) DIC diffusion. The vertical lines indicate the location of the top of the ridge.

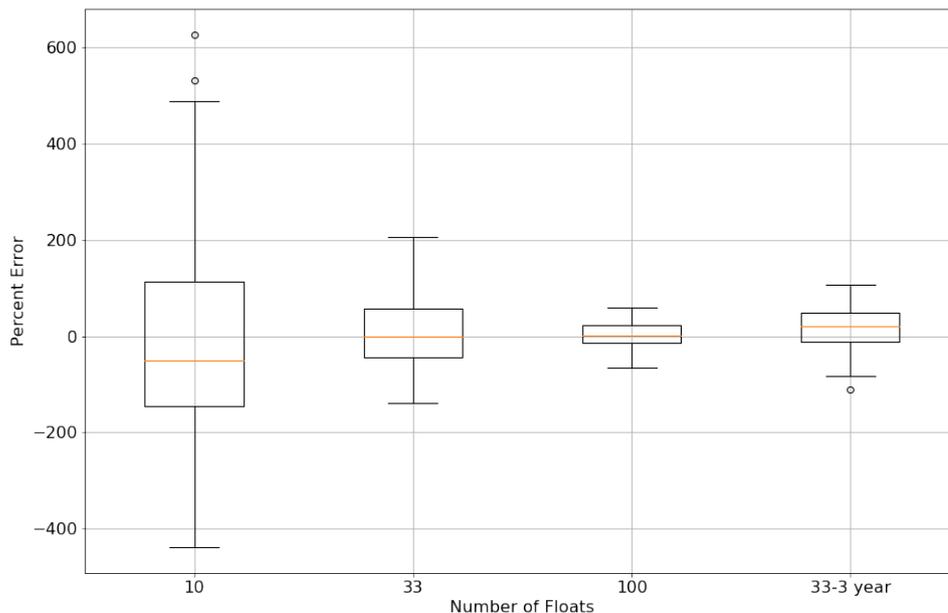


Figure 3. Percent error in the domain-integrated sea-air CO_2 fluxes with Argo-like model sampling for 10, 33, and 100 floats advected for one year and 33 floats advected for 3 years, respectively. The results of 100 trials with randomly initialized floats are shown for each float density. The boxes show the interquartile range and median (orange line) and the whiskers show 1.5 times the interquartile range over the 100 trials. Positive numbers represent anomalous CO_2 outgassing in the float estimate.

163 DIC advection (Figure 2), with horizontal DIC advection playing a key role (SI Figure
 164 4).

165 **3.2 Sampling heterogeneous carbon fluxes**

166 Topography-induced heterogeneity may challenge observation of ocean carbon pro-
 167 cesses. We quantify the ability of autonomous, Lagrangian floats to sample surface ocean
 168 DIC and associated CO_2 fluxes by adding idealized particles to our model domain. These
 169 particles are transported by the model circulation at 1000 m and sample the surface once
 170 every 10 days, mimicing the behavior of Argo floats. Subsampled surface ocean pCO_2
 171 from the simulated floats is mapped to the full model domain, and then the mapped pCO_2
 172 used to calculate CO_2 flux. We test four deployment strategies (1) 10 floats for one year,
 173 (2) 33 floats for one year, (3) 100 floats for one year, and (4) 33 floats for 3 years. For
 174 each float number and duration we select 100 collections of random initial conditions.

175 We calculate the error by subtracting the model truth from the calculated air-sea CO₂
176 fluxes, integrating over the residual and normalizing by the integrated value of the model
177 truth air-sea CO₂ fluxes. As such, our error estimate is fairly conservative; the error would
178 certainly be larger using a square error metric.

179 Our idealized sampling approach reveals substantial biases in the domain-integrated
180 CO₂ flux, as compared to the model truth. With 10 floats, the interquartile range of the
181 air-sea CO₂ flux error is large, from a 113% overestimate to a -146% underestimate, with
182 larger extremes in the upper and lower 25% of the realizations. In this case, the median
183 error (median = -50%, mean = -11%) is an underestimate of the net fluxes. With 33 floats
184 over 1 year the interquartile range is smaller but still quite large – a 57% overestimate
185 to a -45% underestimate (with mean = 2% and median = -1%). With 100 random floats,
186 the error is substantially smaller with an interquartile range of -13% to 23%, and the me-
187 dian (1%) and mean (3%) indicate an overestimate of the carbon flux. When we advect
188 33 floats for 3 years, the error is larger than 100 floats for a single year, with an interquar-
189 tile range of -11% to 48% and a positive flux bias (mean = 18%, median = 20%). Our
190 analysis reveals that the interquartile range of the error of air-sea CO₂ fluxes is quite large
191 when we simulate a float density comparable to the current SOCCOM array (33 floats
192 in a 4000 km sector of the Southern Ocean). Both adding more floats and advecting the
193 floats for 3 years reduces the error. However, even in the absence of interannual variabil-
194 ity, 33 floats advected for 3 years has an increased error range and a positive bias when
195 compared with 100 floats for 1 year.

196 The bias in the idealized float-like sampling of surface carbon arises from the in-
197 fluence of topography on the float trajectories (Figure 4). As an example of the influ-
198 ence of topographically influenced sampling on the calculated air-sea CO₂ fluxes, we show
199 annual-mean fluxes derived from the model (Figure 4a), calculated using the mapped
200 pCO₂ as sampled by 33 floats (Figure 4b; float trajectories in black), and the difference
201 between the model truth and the subsampled fluxes, where blue indicates an underes-
202 timate by the floats and red is an overestimate (Figure 4c). In this example, the floats
203 produce a large underestimate of flux upstream of the ridge due to a lack of sampling
204 in this region (Figure 4c). However, the CO₂ flux is overestimated in other regions (Fig-
205 ure 4c), such the net error is an overestimate of 19%. Particles tend to follow barotropic
206 streamlines as they circumnavigate the Southern Ocean in our model (e.g., Figure 4b).
207 Despite the random initial particle seeding, particles tend to undersample the region up-

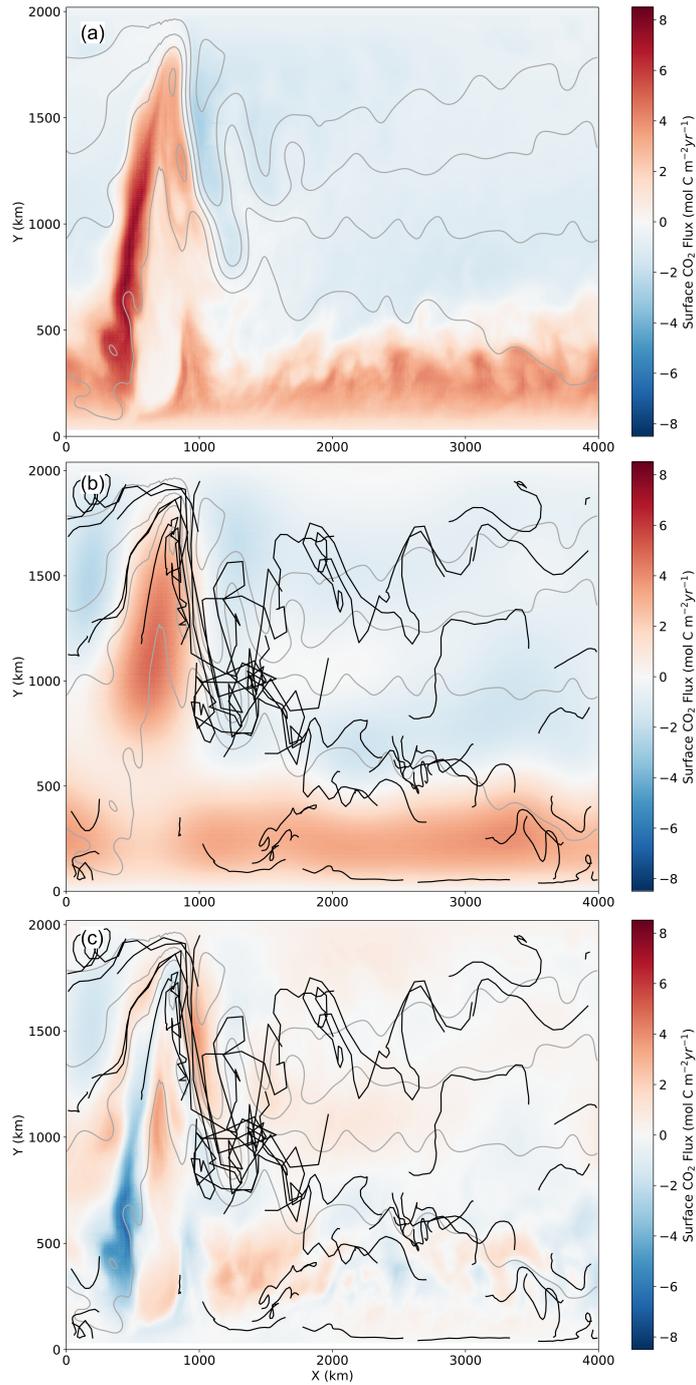


Figure 4. Air-sea CO₂ fluxes over one year derived from the idealized channel model. (a) Modeled fluxes, (b) fluxes as sampled by 33 randomly spaced particles, and (c) the difference between the sub-sampled fluxes and the model truth, with a 19% overestimate of the fluxes. Grey contours indicate barotropic streamlines, while black lines show the tracks of the 33 floats used to generate the images in panels (b) and (c).

208 stream of topography where streamlines are close together (e.g., Figure 4b) and over-
209 sample the region downstream of topography (e.g., Figure 4b) where eddy kinetic en-
210 ergy is at a maximum (Figure 1a).

211 4 Conclusions and Discussion

212 Using an idealized channel model of the Southern Ocean with an undersea ridge,
213 we examine the influence of topography on air-sea CO₂ fluxes. We find intense sea-air
214 CO₂ fluxes and elevated surface ocean DIC upstream of topography, driven by enhanced
215 DIC advection. Due to the nature of the flow near topography, Argo-like particles in our
216 model tend to undersample the region upstream of the ridge and oversample the region
217 downstream of the ridge, leading to biases in domain-integrated CO₂ fluxes.

218 In a previous paper using the same idealized model, Youngs and Flierl (2023) find
219 localized upwelling upstream of the topographic ridge in association with a standing eddy;
220 this localized upwelling is collocated with the region of enhanced CO₂ outgassing reported
221 in this study, suggesting that the standing eddy induced by topography can affect air-
222 sea CO₂ exchange. As water parcels approach the ridge, the flow is deflected northward,
223 which also steepens the isopycnal surfaces and produces a vertical flux of DIC consis-
224 tent with the along-isopycnal vertical tracer flux mechanism described in Freilich and
225 Mahadevan (2019).

226 The largest outgassing is associated with the barotropic effect of topographic fea-
227 tures. Lagrangian floats advected at 1000 m are influenced by this topographic effect.
228 Our results show that Lagrangian particle density is highest in regions with the high-
229 est EKE, consistent with the study of Wang et al. (2020). Yet, our model predicts that
230 the region associated with the highest DIC and thus the largest sea-air CO₂ flux occurs
231 upstream of the ridge, in a region with large gradients in barotropic flow and DIC that
232 tends to be undersampled by Lagrangian particles. Our findings suggest that Lagrangian
233 floats may also undersample topographically induced biogeochemical anomalies (e.g., DIC,
234 oxygen, nitrate).

235 Future efforts in observational network design should consider alternate means to
236 estimate the biogeochemistry of topographically influenced regions. One approach is to
237 use alternative technologies such as gliders (e.g. Dove et al., 2021). This study uses the
238 current standard Gaussian objective mapping technique to map surface ocean pCO₂ and

239 infer air-sea CO₂ fluxes. A complementary approach to confronting the challenges posed
240 by Lagrangian autonomous sampling platforms is developing mapping techniques that
241 account for heterogeneous environments such as techniques that utilize information about
242 correlation length scales (Chamberlain, 2022), and those that use ancillary data such as
243 temperature and salinity to map biogeochemical variables (A. Gray, pers. comm.). Such
244 approaches may improve the sampling error in topographically influenced regions.

245 The idealized model geometry used in this study has enabled mechanistic insights
246 into the drivers of outgassing hotspots at topographic features in the Southern Ocean
247 (Tamsitt et al., 2017; Brady et al., 2021). The insight that barotropic effects have a pri-
248 mary role in driving outgassing hotspots has direct implications for observing system de-
249 sign. Increasing model complexity through more complex and realistic model geometry,
250 improved realism of multiple biogeochemical tracers, finer resolution model configura-
251 tions, and seasonal variability that can improve representation of wind-current interac-
252 tions (Kwak et al., 2021) may enable additional insights about the ways that zonal asym-
253 metry influences the Southern Ocean carbon cycle and the coupling between DIC and
254 other biogeochemical factors in the Southern Ocean.

255 Seafloor topography induces anomalies in both the flow and the surface ocean DIC
256 concentration, leading to sub-optimal sampling of a key region for Southern Ocean CO₂
257 flux. Through the mechanistic insight provided by this study, we suggest that the cur-
258 rent SOCCOM float array has most likely undersampled (rather than oversampled) po-
259 tential areas of CO₂ outgassing in the Southern Ocean, which could further amplify the
260 differences in CO₂ fluxes estimated from SOCCOM floats and those estimated from ship-
261 based observations (Gray et al., 2018; Bushinsky et al., 2019). Topographically influenced
262 regions in the Southern Ocean should be a focus for future biogeochemical observation
263 and modeling programs.

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273 available at <https://www.glodap.info/>.

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