

1 **Interactions between multiple physical particle**
2 **injection pumps in the Southern Ocean**

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9 **Key Points:**

- 10 • Observations from a quasi-Lagrangian float and ocean glider provide insight into
11 physical contributions to the biological pump.
12 • Regimes are identified when mixed layer and eddy subduction pumps are in op-
13 position and work in tandem.
14 • The vertical structure of eddy stirring and transport by submesoscale motions strongly
15 influences the coupling of various biological pumps.

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Abstract

The biological pump, which removes carbon from the surface ocean and regulates atmospheric carbon dioxide, comprises multiple processes that include but extend beyond gravitational settling of organic particles. Contributions to the biological pump that arise from the physical circulation are broadly referred to as physical particle injection pumps; a synthetic view of how these physical pumps interact with each other and other components of the biological pump does not yet exist. In this study, observations from a quasi-Lagrangian float and ocean glider, deployed in the Southern Ocean's subantarctic zone for one month during the spring bloom, offer insight into daily-to-monthly fluctuations in the mixed layer pump and the eddy subduction pump. Estimated independently, each mechanism contributes intermittent export fluxes on the order of several hundreds milligrams of particulate organic carbon (POC) per day. The float and the glider produce similar estimates of the mixed layer pump, with sustained weekly periods of export fluxes with a magnitude of ~ 400 mg POC $\text{m}^{-2} \text{day}^{-1}$. Export fluxes from the eddy subduction pump, based on a mixed layer instability scaling, occasionally exceed 500 mg POC $\text{m}^{-2} \text{day}^{-1}$, with some periods having strong inferred vertical velocities and others having enhanced isopycnal slopes. Regimes occur when a summation of the two pump estimates may misrepresent the total physical carbon flux. Disentangling contributions from different physical pump mechanisms from sparse data will remain challenging. Insight into how mesoscale stirring and submesoscale velocities set the vertical structure of POC concentrations is identified as a key target to reduce uncertainty in global carbon export fluxes.

Plain Language Summary

The ocean influences the global carbon cycle by transferring carbon from the surface into the deep ocean, where it is sequestered from exchange with the atmosphere for periods of decades to millennia. Marine organisms enhance this downward transfer by fixing dissolved CO_2 into organic matter during photosynthesis (at the ocean surface) and removing the carbon through sinking of organic matter or via consumption and defecation by higher trophic levels. These processes are collectively referred to as the biological pump. Processes controlled by the ocean circulation, termed "physical" processes, can also influence the vertical transfer of fixed carbon to the ocean interior. Various proposed physical carbon pump mechanisms have been assessed independently without considering potential interactions between them. Here, we use data sets collected from different autonomous vehicles deployed in the Southern Ocean to provide a first look at interactions between two physical carbon pumps: the mixed layer and eddy subduction pumps. These pumps make a significant contribution to total carbon export, but there are times when they work in tandem and times when they work in opposition. The vertical structure of particulate organic carbon is identified as a key target for future observations to better constrain the biological pump.

1 Introduction: Physically-induced particle pumps as a component of the biological pump

A small fraction of the ocean’s volume, limited to the well-mixed ocean surface boundary layer, or the mixed layer, directly exchanges carbon dioxide (CO_2) with the atmosphere. Any process that transports carbon, in particulate or dissolved form, across the base of the mixed layer removes it from further exchange with the atmosphere for timescales spanning months to millenia. This timescale is primarily determined by the penetration depth of the carbon export before remineralization occurs (Yamanaka & Tajika, 1996; Kwon et al., 2009). The ocean’s large scale, overturning circulation is associated with the solubility pump, in which cold waters, rich in dissolved inorganic carbon (DIC) are preferentially carried to depth in polar regions and exported to lower latitudes (Sarmiento & Gruber, 2006). In contrast, biological processes that fix carbon as organic matter in the surface ocean and export the carbon as particulate organic carbon (POC) via gravitational sinking, the biological gravitational pump (BGP), is a global process (Siegel et al., 2023). Observations collected over the past decade provide convincing evidence that, at least regionally, physical processes that advect and stir surface properties to depth, known as physically-induced particle injection pumps (PPIPs), make a non-negligible contribution to the biological pump (Lévy et al., 2012; Omand et al., 2015; Stukel & Ducklow, 2017; Boyd et al., 2019). Carbon export due to PPIPs depends on both ecosystem organization, *e.g.* surface chlorophyll concentration and community composition, and physical properties, *e.g.* mixed layer depth and front intensity, which vary from sub-kilometer to many tens of kilometers spatial scales (Mahadevan, 2016; Lévy et al., 2018; Taylor, 2018; Ruiz et al., 2019). PPIPs are typically more spatially heterogeneous than the BGP or the solubility pump, and quantification of carbon fluxes due to various PPIPs remains immature.

The eddy subduction pump (ESP) refers to processes that enhance vertical velocities and vertical transport of dissolved and organic carbon across the base of the mixed layer. Most of the energy in the global ocean resides in geostrophically-balanced, mesoscale flows that are predominantly horizontal. Geostrophic balance begins to break down at scales roughly smaller than 10 km, the submesoscale dynamical regime, at which vertical velocities strengthen through ageostrophic motions. A number of physical processes can elevate vertical velocities or turbulent mixing at the ocean’s submesoscale as described in reviews by McWilliams (2016) and Taylor and Thompson (2023). Submesoscale advective fluxes are generally localized to small-scale fronts and arise from a tendency for ageostrophic cross-frontal vertical velocities to reduce the strength of the front. Most previous studies of the submesoscale ESP have focused on the role of mixed layer baroclinic instability, in which the magnitude of vertical velocities and export fluxes can be related to observable surface properties, such as mixed layer depth (MLD) and the magnitude of horizontal density gradients in the ocean’s surface boundary layer (Omand et al., 2015), with some caveats. In particular, the theory underpinning the parameterization assumes that vertical velocities peaks at the center of the mixed layer and decay to zero at the mixed layer base, implying negligible carbon export out of the mixed layer. Yet, submesoscale motions may extend deeper into the water column (below the mixed layer) as the stratification at the base of the mixed layer weakens (Callies et al., 2016; Erickson & Thompson, 2018). Nevertheless, mixed layer baroclinic instability has provided a useful approach for estimating the relative importance of submesoscale vertical velocities on a global scale. Omand et al. (2015) estimate that for large swaths of the ocean, notably the North Atlantic and the Southern Ocean’s Subtropical Frontal Zone, eddy subduction may account for 25%-50% of springtime carbon export.

While vertical carbon fluxes are enhanced at submesoscale fronts (Ruiz et al., 2009), mesoscale eddies contribute to the generation and distribution of these fronts, especially in areas of strong strain at eddy peripheries that support frontogenesis (Brannigan, 2016; Thomas et al., 2013; Su et al., 2020; Siegelman et al., 2020; Freilich & Mahadevan, 2021).

107 Mesoscale stirring properties are spatially heterogeneous in the ACC, which modifies tracer
108 variance on isopycnal surfaces (Dove et al., 2021) and exchange between the mixed layer
109 and ocean interior (Brady et al., 2021; Dove et al., 2022). Llort et al. (2018) found ev-
110 idence of largely unmodified surface waters, identified by low apparent oxygen utiliza-
111 tion (AOU) values, hundreds of meters below the mixed layer using biogeochemical-Argo
112 floats. These anomalies were localized to regions of high eddy kinetic energy (EKE) found
113 in the lee of major Southern Ocean topographic features, but were only identified in \sim
114 1% of all profiles. Siegelman et al. (2020) used high-resolution measurements collected
115 from an instrumented elephant seal to suggest that mesoscale frontogenesis enhances ver-
116 tical velocities well below the base of the mixed layer. Finally, Siegelman et al. (2020)
117 and Dove et al. (2022) showed the utility of the Lagrangian-based diagnostic, finite size
118 Lyapunov exponents (FSLEs, section 2.4; d’Ovidio et al. (2004)), for mapping spatial
119 variations in small-scale subduction in the Southern Ocean with exchange again elevated
120 in regions of strong mesoscale strain. These studies highlight the important coupling be-
121 tween mesoscale and submesoscale motions; the impact of this coupling on the biolog-
122 ical pump remains underexplored.

123 Mesoscale motions and coherent mesoscale eddies not only modulate the magni-
124 tude and distribution of the upper ocean submesoscale velocity field, but they also con-
125 tribute to the delivery of surface tracers and particulate matter to depth. Recent stud-
126 ies suggest that small-scale fronts are the sites where surface properties are imprinted
127 on density surfaces immediately below the base of the mixed layer. This establishes a
128 concentration gradient on that isopycnal that is subsequently stirred out by mesoscale
129 processes (Balwada et al., 2018; Ruiz et al., 2019; Freilich & Mahadevan, 2021). The edges
130 of mesoscale eddies are consistently identified as key locations of enhanced surface-interior
131 exchange, enhanced isopycnal tracer gradients, and intensified isopycnal tilting. Mesoscale
132 eddies may also modify biomass and nutrient availability through Ekman pumping and
133 isopycnal displacement, further influencing carbon export (Rohr et al., 2020a, 2020b).
134 Finally, the influence of mesoscale motions can extend beyond passive tracers and influ-
135 ence biology, for example through controls on nutrient transport (Patel et al., 2020) and
136 the structure of surface chlorophyll (Cornec et al., 2021). Of particular relevance here,
137 Penna et al. (2022) report finescale interleaving of distinct mesopelagic micronekton com-
138 munities along the periphery of a mesoscale eddy in the Southern Ocean Subantarctic
139 Zone, close to the location of our present study.

140 The mixed layer pump (MLP) is a PPIP traditionally associated with one-dimensional
141 (vertical) processes that influence the seasonal cycle of MLD. Export via the mixed layer
142 pump over an annual timescale is associated with the accumulation of POC in the sur-
143 face mixed layer during spring and summer growth seasons, a redistribution of this POC
144 to greater depths as the mixed layer deepens in winter, and a final “export” or detrain-
145 ment of carbon as the ML shoals in the spring (Gardner et al., 1995). However, the MLP
146 may act on shorter timescales, spanning months to weeks, or even days, the latter as-
147 sociated with abrupt restratification events caused by submesoscale mixed layer insta-
148 bilities (Thompson et al., 2016). Dall’Olmo et al. (2016) combined observations from Argo
149 floats and remote sensing data to estimate the magnitude of the seasonal MLP and found
150 that in high-latitude regions, where large fluctuations in MLD are common over an an-
151 nual cycle, up to a quarter of the carbon flux to the mesopelagic zone occurs through
152 the MLP; in localized regions the MLP can account for 100% of the carbon export. Higher
153 frequency sampling from ocean gliders, as compared to Argo floats, provided the oppor-
154 tunity to estimate daily POC flux estimates from the MLP at daily temporal resolution
155 in the North Atlantic and revealed a qualitatively similar importance of the MLP (Bol
156 et al., 2018). Specifically, small-particle POC transfer across the base of the mixed layer
157 was most efficient in winter and early spring, contributing between 5% and 25% of the
158 total export flux. Daily estimates of POC export via the MLP produced large fluxes re-
159 lated to both export and entrainment, with a small residual emerging at monthly timescales.

160 The amplitude was particularly large in the January to May time period, associated with
 161 high frequency fluctuations in the MLD at this location (Damerell et al., 2020).

162 The physical and biological mechanisms that carry POC across the base of the mixed
 163 layer in the Southern Ocean and modify tracer properties on interior density surfaces
 164 has an out-sized impact on the global marine carbon cycle. Tracers in the Southern Ocean
 165 undergo strong stirring due to the generally high levels of EKE while also residing in re-
 166 gions of relatively weak vertical surface stratification, which may precondition the bi-
 167 ological pump to have substantial contributions to export from the ESP and the MLP
 168 (Omand et al., 2015; Gille et al., 2022; Lacour et al., 2023). Additionally, the seasonal
 169 cycle of sea ice melt induces large-scale surface density gradients cascade to smaller scales
 170 via mesoscale stirring, producing in small-scale density fronts susceptible to hydrody-
 171 namic instabilities and enhanced vertical velocities (Giddy et al., 2021, 2022). The preva-
 172 lence of these submesoscale processes are not uniformly distributed over the Southern
 173 Ocean, but are concentrated in regions of high EKE downstream of major topographic
 174 features (Gille & Kelly, 1996), which are also hotspots of subduction (Dove et al., 2023).
 175 The increased deployment of biogeochemical-Argo floats, in particular through the SOC-
 176 COM project (Bushinsky et al., 2019), has improved understanding of the large-scale dis-
 177 tribution of biogeochemical properties in the Southern Ocean, and a few studies have
 178 shown how these floats can help to capture regional and smaller-scale processes that shape
 179 these distributions (Llort et al., 2018) and their interannual variability (Lacour et al.,
 180 2023).

181 In this study, we provide estimates of carbon export PPIPs in the Southern Ocean
 182 using observations from two different autonomous platforms: an ocean glider and a quasi-
 183 Lagrangian float (section 2). These observations were collected as part of the SOLACE
 184 (Southern Ocean Large Areal Carbon Export) experiment, which is a contribution to
 185 the JETZON (Joint Exploration of the Twilight Zone Ocean Network) effort to assess
 186 export processes in various regions of the global ocean. During the deployment, the ocean
 187 glider was piloted to provide spatial context of mesoscale and submesoscale physical and
 188 biogeochemical tracer variability surrounding the water-mass tracking float, with a goal
 189 of attributing particle export to various carbon pump mechanisms. The existing liter-
 190 ature on PPIPs tends to focus on a single process, *i.e. either* the ESP or MLP, with lit-
 191 tle previous consideration of how carbon pumps interact. We find that classification of
 192 PPIPs requires analysis of combined data sets, and consideration of tracer vertical struc-
 193 ture, and even then can remain difficult to separate. Here, we rely on existing approaches
 194 and parameterizations to estimate PPIP magnitudes (section 3), and focus on analysis
 195 of the observational data (section 4) to offer insight in accounting for PPIPs in global
 196 estimates of the biological pump (section 5).

197 **2 Data and Methods: a float-glider observing pair**

198 **2.1 Study region and conditions**

199 The Southern Ocean Time Series (SOTS) observatory, established in 1997, is lo-
 200 cated at approximately 47°S, 142°E, 500 km southwest of Tasmania, Australia, in 4500 m
 201 water depth (Figure 1). SOTS is located north of a region where the Subantarctic and
 202 Polar fronts (SAF and PF) typically merge along the eastward flowing ACC. These fronts
 203 form the southern boundary of a small anticyclonic gyre, which encompasses SOTS, and
 204 is bounded to the north by subtropical waters flowing westward along the East Australian
 205 Current extension (Herraiz-Borreguero & Rintoul, 2011; Trull et al., 2001). Sediment traps
 206 moored at SOTS show a long history of hydrographic and biogeochemical properties broadly
 207 characteristic of Subantarctic waters between 90°E and 145°E (Trull et al., 2001; Shad-
 208 wick et al., 2015). The proximity of the SOTS region to the SAF and PF means that
 209 this area is frequently populated by coherent mesoscale eddies, which are associated with
 210 anomalies in surface temperature and chlorophyll concentrations (Figure 1a).

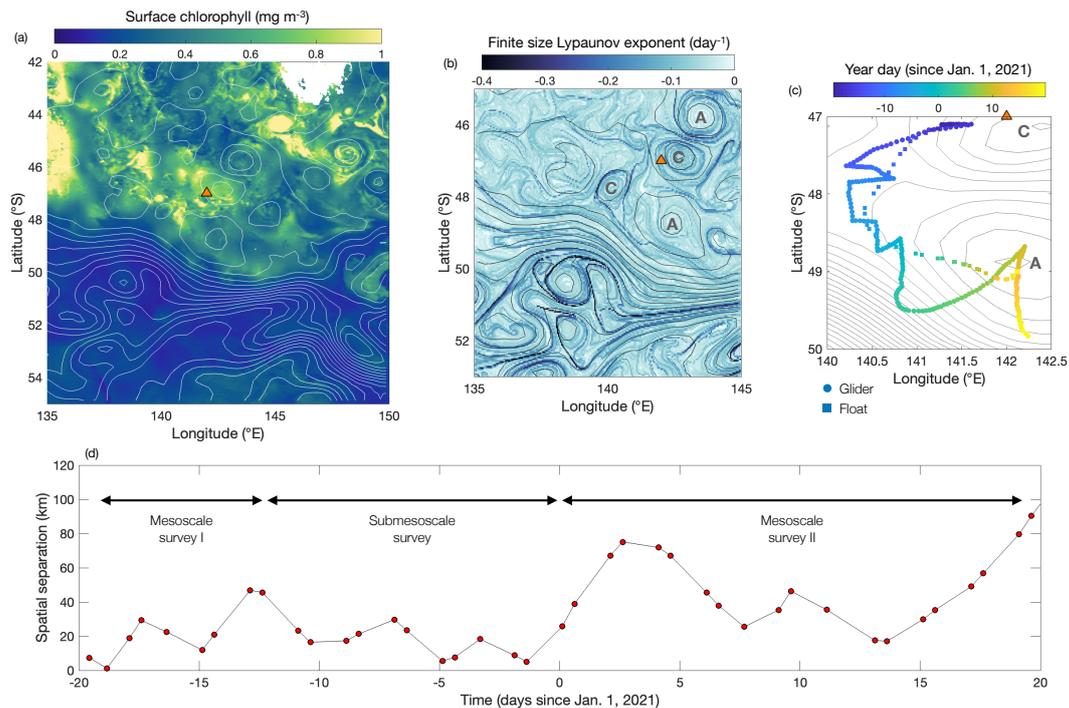


Figure 1. Overview of the SOLACE study region. (a) Snapshot of surface chlorophyll concentration (color) and sea surface height (SSH) contours (white) on December 21, 2020. (b) Satellite-derived finite size Lyapunov exponents (FSLE; section 2.4) and SSH contours (black) for December 21, 2020, highlighting regions of strong strain (negative FSLE values) at the periphery of the mesoscale eddies. The letters “A” and “C” refer to eddies with anticyclonic and cyclonic vorticity, respectively. (c) Enlarged view of SSH contours (black) and the position of the float (squares) and glider (circles), colored by year day; the SSH contours in this panel are a time average between 15 December, 2020 and 15 January, 2021. (d) Spatial separation between the float and glider (km) during the initial stage of the deployment; the sampling was divided into two longer glider mesoscale surveys and a period of shorter transects where the glider sampled within ~ 20 km of the float. The time axis represents the number of days since January 1, 2021; observations from December 2020 are reported as negative days. In the upper panels, the position of the Southern Ocean Time Series (SOTS) long-term mooring is given by the orange triangle, for reference.

211 The SOLACE project seeks to quantify the contribution of both physically- and
 212 biologically-mediated particle injection pumps to the total export of carbon. The project
 213 combines ship-based, autonomous, and remote sensing observations. Here, we focus on
 214 data collected from two autonomous platforms, a quasi-Lagrangian float equipped with
 215 a Underwater Vision Profiler 6 (UVP6; WMO 5906623) and a Seaglider (SG674). These
 216 platforms were deployed on December 12, 2020 at 47.1°S , 141.3°E from the *R/V Investigator*.
 217 Altimetry-derived sea surface height (SSH, section 2.4) indicates that in mid-
 218 to-late December the SOLACE study region contained two cyclonic circulation features,
 219 centered at 47.8°S , 140.3°E and 47.0°S , 143.0°E and an anticyclonic circulation feature,
 220 centered at 48.8°S , 142.0°E . These cyclonic and anticyclonic eddies were associated with
 221 positive and negative surface chlorophyll anomalies, respectively (Figure 1a). The float
 222 and glider sampled in and around these eddies while remaining north of the ACC’s SAF.
 223 The float’s trajectory largely followed the geostrophic circulation inferred from the SSH

224 contours. Following deployment, the float drifted to the southwest for a period of ~ 2.5 weeks,
 225 until reaching the northern edge of the SAF, after which it drifted eastward (Figure 1c).
 226 In early January 2021, the float drifted into the core of the southern anticyclone, where
 227 it remained relatively stationary over a period of roughly 2 weeks before drifting back
 228 towards the northern edge of the eddy. The glider closely tracked the float position un-
 229 til mid-January at which time the salinity sensor on the glider became fouled; we thus
 230 focus on the period between mid-December and mid-January. Over this time, the glider
 231 completed over 200 V-shaped dives to 1000 m, producing more than 400 profiles, account-
 232 ing for up- and down-casts. Each dive had a duration between 3 and 5 hr and covered
 233 a horizontal distance between 2 and 4 km. During this same period, the float completed
 234 37 profiles.

235 The glider was deployed to provide both mesoscale and submesoscale context around
 236 the float. The glider sampling pattern involved multiple transects spanning roughly 10
 237 to 100 km, and where possible, aligned these transects perpendicular to the float's tra-
 238 jectory. The glider survey can roughly be divided into three periods: (i) a mesoscale tran-
 239 sect that assessed the density structure of the two cyclonic eddies located in the north-
 240 ern extent of the study domain; (ii) a period of shorter, rapid transects that crossed the
 241 float's trajectory with near coincident occupations of the same location (typically within
 242 less than a day); (iii) a second mesoscale transect that sampled both the SAF and the
 243 southern boundary of the anticyclone. These different periods are indicated in Figure
 244 1d, which shows the spatial separation between the glider and float during the deploy-
 245 ment. The time associated with each platform is provided as time in number of days since
 246 January 1, 2021; observations from December 2020 are reported as negative days.

247 We make use of two derived properties, spice and AOU. Spice is defined as the com-
 248 bination of temperature and salinity that is locally orthogonal to isolines of potential den-
 249 sity, such that spice variance is fully density compensated (Flament, 2002). AOU is de-
 250 fined as the difference between oxygen solubility, determined from temperature and salin-
 251 ity, and the measured oxygen concentration, $\text{AOU} \equiv [O_2]_{\text{sat.}} - [O_2]_{\text{obs.}}$ (Ito et al., 2004).

252 2.2 Float data and processing

253 The float was a CTS5 Provor float developed by Nke Instrumentation. It carried
 254 a Sea-Bird SBE41 sensor measuring conductivity, temperature and pressure, an Aanderaa
 255 Oxygen Optode, and a WET Labs ECO Triplet measuring induced chlorophyll-a and
 256 Colored Dissolved Organic Matter (CDOM) fluorescence, and the volume scattering func-
 257 tion at 700 nm and angle of 124°S . The volume scattering function includes scattering
 258 signal from pure seawater and particulate scattering (Zhang et al., 2009; Vaillancourt
 259 et al., 2004). The scattering by seawater was calculated using a function described in Zhang
 260 et al. (2009) and subtracted from the volume scattering function. The resulting partic-
 261 ulate volume scattering function was converted into particulate optical backscattering
 262 coefficient bbp (Bol et al., 2018; Briggs et al., 2011). Finally, following Briggs et al. (2011),
 263 a despiked backscatter data product is also produced using a seven-point min-
 264 imum filter followed by a seven-point maximum filter to remove spikes, which often oc-
 265 cur in profiles of optical backscatter due to aggregate material. The CTD and trajec-
 266 tory data were quality-controlled using the standard Argo protocol (Wong et al., 2020).
 267 Oxygen data was calibrated to the ship-based CTD cast performed less than an hour af-
 268 ter float and glider deployment.

269 The float mission included pairs of day and night profiles every 2 days from 1000 m
 270 depth to the surface at noon and from 500 m at night, with a parking depth of 1000 m.
 271 Occasionally, the float profiled down to 2000 m for the calibration of sensors not used
 272 in this study. CTD and bbp data were acquired at a vertical resolution of 0.1 m and 1 m,
 273 respectively. Oxygen data were acquired at a vertical resolution ranging from 0.1 m to
 274 10 m depending on the depth layer.

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2.3 Glider data and processing

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The glider carried an unpumped CTD (CT-Sail) sensor measuring conductivity (salinity), temperature, and pressure; an Aanderaa oxygen optode; and a WET Labs ECO puck that measured induced fluorescence and optical backscatter. The salinity, temperature, and dissolved oxygen measurements from the glider sensors were calibrated to ship-based hydrographic data collected at the deployment location during the SOLACE cruise. The submesoscale phase of the glider sampling, the period of rapid crossings of the float trajectory (Figure 1b), provided additional opportunities to confirm the calibration between the two sensors. These revealed no significant sensor drift in temperature, salinity or oxygen. To conserve battery power, optical measurements were collected down to 500 m with occasional dives down to 1 km to determine a background signal. Optical backscatter data on the gliders were measured at two wavelengths: 470 and 700 nm. The glider backscatter and oxygen data were processed in the same way as the relevant float data. At times throughout the deployment when the float and glider were within 5 km of the other, parameters from the two platforms were cross-checked but no further inter-calibration was deemed necessary.

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In addition to using the ungridded, processed glider data, measurements from the glider were objectively mapped onto a regular grid with 10-m depth resolution along the vertical axis and 500-m distance resolution along the horizontal axis, using a Gaussian weighting function with a vertical scale of 20 m and a horizontal scale of 5000 m, *cf.* (Viglione et al., 2018). A visual comparison of the raw data to the objectively mapped data set revealed no significant biases due to this choice of resolution. The time associated with each glider position was interpolated to the horizontal distance grid.

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2.4 Altimetry-derived parameters

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Daily estimates of SSH relative to the geoid for the duration of the SOLACE field program were obtained from the L4-gridded satellite altimetry product provided by Copernicus Marine Services. This includes the following data products: absolute dynamic topography (ADT), sea level anomaly (SLA), total geostrophic velocity, and eddy geostrophic velocity. The latter two fields are associated with the gradients of ADT and SLA fields, respectively.

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Satellite altimetry-derived surface geostrophic velocities were used to calculate the Okubo-Weiss (OW) parameter each day during December 2020 and January 2021 (Figure 2b) and to interpolate these values in time and space to the position of the float and glider (Figure 2c). The OW parameters is given by:

$$\text{OW}_g \equiv s_n^2 + s_s^2 - \zeta^2, \quad (1)$$

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where the subscript ‘g’ indicates that OW is estimated from geostrophic velocities, and terms on the right hand side refer to the normal strain $s_n = u_x - v_y$, the shear strain $u_y + v_x$, and the vertical relative vorticity $v_x - u_y$. Regions where $\text{OW} > 0$ and $\text{OW} < 0$ are dominated by straining or vortical motions, respectively. The altimetry data, provided on a 25-km grid but resolving structures at larger, $O(100 \text{ km})$ scales, underestimates the magnitude of vorticity and strain, which is enhanced at meso- and submesoscales. Nevertheless, during the SOLACE campaign, the float broadly occupied a region dominated by straining motions, which are more conducive to frontogenesis and the generation of submesoscale surface fronts and eddies that enhance vertical velocities. The glider sampled both vorticity- and strain-dominated regions, in particular during the two mesoscale survey period (Figure 1c).

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Finite-size Lyapunov exponents (FSLEs) describe the orientation and timescale of strain fields by quantifying stretching and compression (d’Ovidio et al., 2004). They are a Lagrangian diagnostic, and for a given flow field are defined as the separation growth

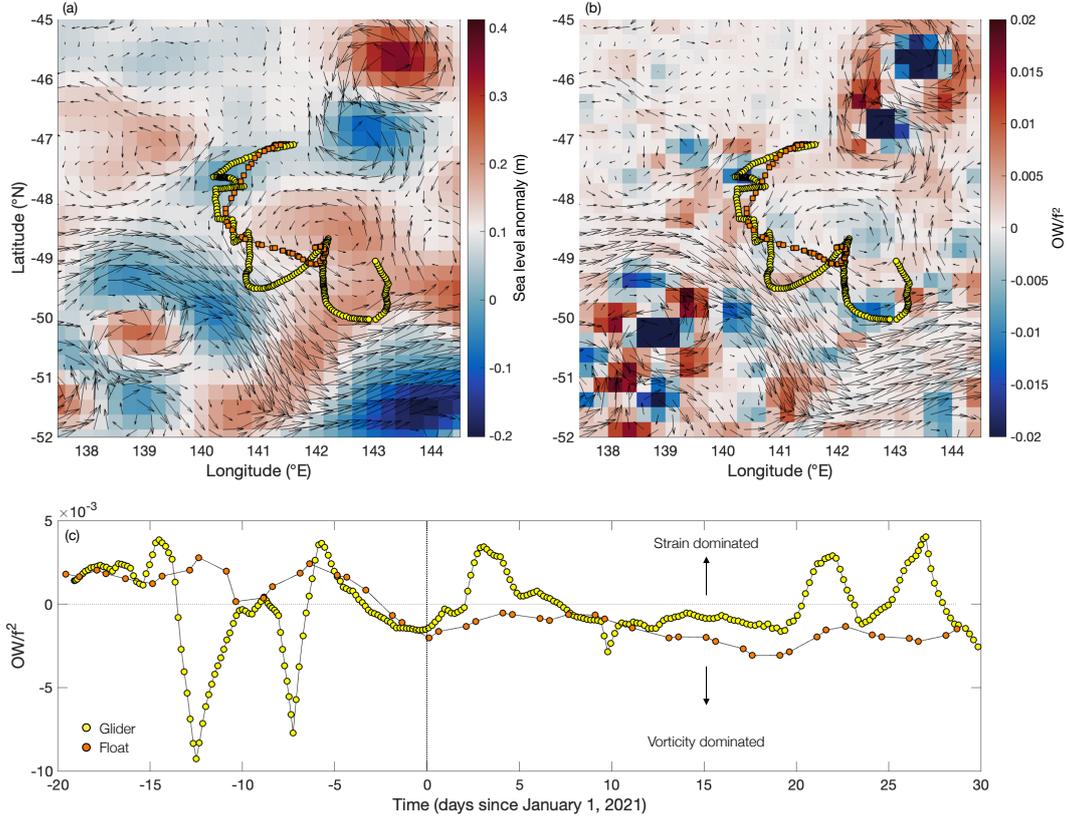


Figure 2. (a) Snapshot of the glider and float positions overlaid on sea level height anomaly (SLA, m, color) and surface geostrophic velocity (arrows) from 1 January, 2021. Coherent regions of cyclonic and anti-cyclonic vorticity are indicated by negative and positive SLA values, respectively. (b) Snapshot of the non-dimensionalized Okubo-Weiss parameter (1) (color), where f^2 is the Coriolis frequency squared, and surface geostrophic velocity (arrows) from 1 January, 2021. Negative and positive values indicate regions dominated by vorticity and strain, respectively. (c) Time series of the OW parameter, normalized by f^2 , interpolated in space and time to the position of the glider (yellow) and float (orange).

323 rate for seeded particle pairs. Here, FSLE estimates were computed from satellite-derived
 324 geostrophic velocities and provided by AVISO+.

325 Links to altimetry data products are provided in section 7.

326 2.5 Surface fluxes

327 Estimates of surface forcing fields for the region preceding, during, and after the
 328 study period were obtained from the SOTS surface mooring, which has a meteorological
 329 sensor. The data is publicly available through the Australian Ocean Data Network
 330 (AODN). Net surface heat flux was calculated as the sum of the shortwave, longwave,
 331 sensible, and latent heat flux estimates. Surface wind stress magnitude is also provided
 332 from the SOTS mooring time series. The zonal and meridional components of the wind
 333 speed, located 10 m above the surface, is calculated from the provided wind speed mag-
 334 nitude and the wind speed direction. The winds are predominantly westerly at the SOTS
 335 location (Figure 3b).

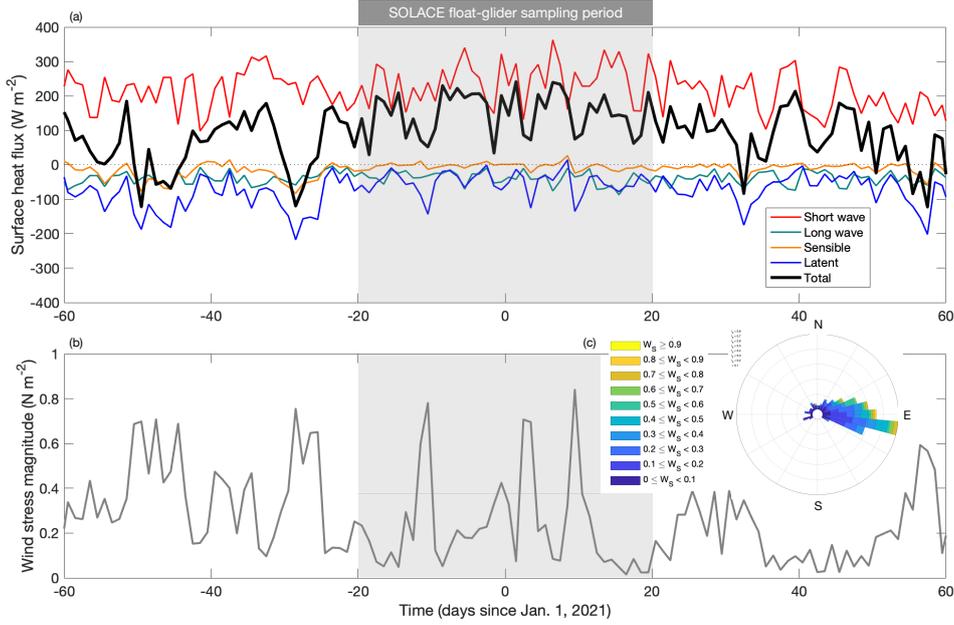


Figure 3. Surface forcing during the SOLACE field experiment as recorded at the SOTS mooring. (a) Contributions to the daily-mean ocean surface heat flux (W m^{-2}): incoming short-wave radiation (red), outgoing longwave radiation (blue), sensible heat flux (orange), latent heat flux (green), and the total (black). (b) Wind stress magnitude derived from surface winds 10 m above the surface and (c) windrose plot showing wind speed and direction. In the windrose plot, direction shows wind speed orientation while the legend indicates wind speed magnitude.

3 Estimation of PPIPs

The main focus of this study is the time series of the ESP and the MLP as estimated from float and glider data. Ocean gliders have proven effective at resolving sub-mesoscale fronts that can be localized sites of enhanced vertical fluxes. Here, we follow the approach outlined in Omand *et al.* (2015) to estimate the magnitude of the ESP, noting that this formulation only accounts for motions related to mixed layer baroclinic instability. The assumption is that a relationship for the vertical export flux of POC by the ESP, $\langle w' \text{POC}' \rangle$, can be linked to submesoscale upper ocean buoyancy fluxes related to near-surface restratification:

$$F_{\text{ESP}} \equiv \langle w' \text{POC}' \rangle \sim \langle w' b' \rangle \left\langle \frac{\partial \text{POC}}{\partial z} \right\rangle \langle N^2 \rangle^{-1}, \quad (2)$$

where $N^2 \equiv -g\rho_0^{-1} (\partial\sigma_0/\partial z) \approx 2 \times 10^{-5} \text{ s}^{-2}$. The final step connects the vertical buoyancy flux $\langle w' b' \rangle$ to a parameterized estimate of the upper ocean eddy overturning streamfunction ψ_e , such that $\langle w' b' \rangle \sim -(\partial\psi_e/\partial x) b' \approx \psi_e M^2$. Estimating the vertical POC gradient as $[\text{POC}]/h_m$, where $[\text{POC}]$ is the mean mixed layer POC concentration and h_m is the MLD, gives

$$F_{\text{ESP}}(t) \sim \psi_e \Gamma [\text{POC}] h_m^{-1}, \quad (3)$$

where Γ is the isopycnal slope, $\Gamma = M^2/N^2$, where $M^2 \equiv g\rho_0^{-1} (\partial\sigma_0/\partial x)$, and ψ_e is estimated using the Fox-Kemper *et al.* (2008) parameterization $\psi_e = 0.08 M^2 h_m^2 f^{-1}$. Below, M^2 is calculated from glider data with x corresponding to distance along the trajectory. The estimated F_{ESP} (3) varies quadratically with lateral density gradient M^2 ,

and linearly with h_m and POC concentration, but varies inversely with mixed layer stratification N^2 .

The export flux of carbon related to the mixed layer pump is estimated using upper ocean estimates of MLD and optical backscatter following the approach in Dall’Olmo et al. (2016) and Bol et al. (2018). To determine a time series of mixed layer POC concentration, we use the calibration presented in Schallenberg *et al.* (2016)

$$[\text{POC}] = 37,601 \times \text{bbp}_{700} + 4.95 \text{ mg m}^{-3}, \quad (4)$$

where bbp_{700} is the 700 nm optical backscatter value. This calibration uses observations collected during the SAZ-Sense voyage in 2007 and the SOTS voyage in 2018. The POC data from the SAZ-Sense voyage was first calibrated against the beam transmissometer, which was subsequently calibrated against bbp using data from the SOTS voyage in March 2018, when observations from both a beam transmissometer and a backscatter sensor were available on CTD casts.

Following Bol et al. (2018), we treat each observation of the mixed layer from the float and glider as part of the time series and do not explicitly account for spatial variations in the flux estimates. The limitations of this approach are discussed in section 5. The mixed layer pump is estimated by

$$F_{\text{MLP}}(t) = -\frac{1}{\Delta t} \int_{h_m(t-\Delta t)}^{h_m(t)} \overline{[\text{POC}]}(z, t) dz, \quad (5)$$

where $\overline{[\text{POC}]}$ is the mean mixed layer POC concentration between times t and $t-\Delta t$. As defined, a reduction in the MLD ($h_m(t) < h_m(t-\Delta t)$) results in a positive export flux as POC is assumed to be left behind; as the mixed layer deepens, POC is re-entrained and $F_{\text{MLP}} < 0$. The export flux from the mixed layer pump is binned using daily, and 5-day intervals.

4 Results

4.1 Characterization of tracer variability from the float and glider

The combination of float- and glider-based measurements offers regional perspectives on mixed layer-interior exchange that are both approximately Lagrangian (float) and Eulerian (glider). The \sim daily updates of the float position enabled tight sampling between the float and glider, with the latter providing spatial context around the float and resolution of physical processes with higher temporal scales.

Differences in the float and glider tracer variability arise from two aspects of the sampling: (i) the glider resolves variations over shorter spatial and temporal scales and (ii) the float is largely sampling in a Lagrangian framework. Throughout the upper 500 m, the glider data exhibits greater tracer variability on pressure surfaces (Figure 4). Much of the physical variability is compensated, with large variations in temperature and salinity that is not expressed in density. These variations occur over characteristic scales of $O(1 \text{ day}, 10 \text{ km})$, consistent with filamentary, submesoscale anomalies and are therefore not resolved by the float. The glider reveals the sharp nature of these features. For instance, a positive temperature anomaly extending to 400 m depth, that occurs at day -16 (Figure 4a,b) is recorded by both platforms, but is resolved over multiple dives in the glider. The glider also samples mesoscale anomalies in temperature and salinity linked to large vertical displacements in density surfaces (*e.g.*, day -15 to -13 and 0 to 5). These features are associated with the glider crossing coherent mesoscale eddies and the SAF. The temporal evolution of MLD is broadly similar in the float and glider (black curves in Figure 4e,f). MLD derived from the glider varies on time scales shorter than the float sampling period, which may arise from both submesoscale processes that intermittently cause the mixed layer to shoal or deepen (Thompson et al., 2016; Nicholson et al., 2022),

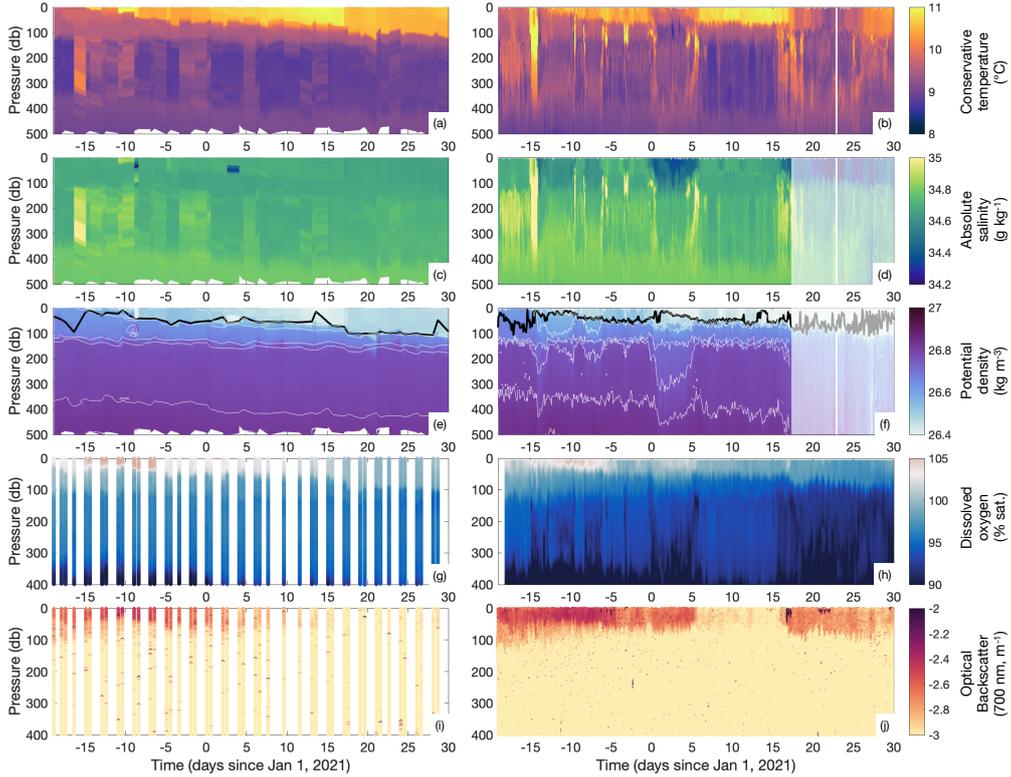


Figure 4. Hydrographic and biogeochemical properties along the trajectory of the float (left) and glider (right). The panels show (a,b) conservative temperature ($^{\circ}\text{C}$), (c,d) absolute salinity (g kg^{-1}), (e,f) potential density σ_0 (kg m^{-3}), (g,h) dissolved oxygen (% saturation), and (i,j) \log_{10} of 700 nm optical backscatter (m^{-1}). The solid black line in panels (e,f) show the mixed layer depth and white contours show density surfaces 26.5, 26.6, 26.7, 26.75, 26.8, 26.85, 26.9 and 26.95 kg m^{-3} . The high frequency noise that occurs from days 18-30 in (d) and (f) are due to a fouling of the salinity sensor; these data are not considered further. The different plotting style in panels (g) and (i) is to indicate that, on the float, the biogeochemical properties are sampled at a lower vertical resolution than the physical properties.

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but may also reflect spatial variability in MLD at scales less than 10 km. The glider’s sampling of small-scale spatial variability in hydrographic properties provides support that the float’s trajectory sampled a distinct water mass during the first month of deployment, despite its vertical profiling and the potential for vertical shear in the water column.

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The distribution of temperature and salinity properties are also more compact in the float, as compared to the glider (Figure 5). Differences between Θ - S_A properties between the float and glider largely occur in the upper ocean, in density layers, $\sigma_0 < 26.7 \text{ kg m}^{-3}$. The float records a near-surface layer, roughly the upper 100 m, where salinity values are uniform and stratification is dominated by temperature; this overlays a second layer, 100-400 m, with compensated temperature-salinity variations; the base of this layer is associated with the wintertime mixed layer depth (Rintoul & Trull, 2001). Deeper density classes $\sigma_0 > 26.8 \text{ kg m}^{-3}$ have similar distributions between float and glider and become progressively colder and fresher with depth, reaching down into AAIW water properties (Trull et al., 2001). The glider samples a broader range of near-surface salinity val-

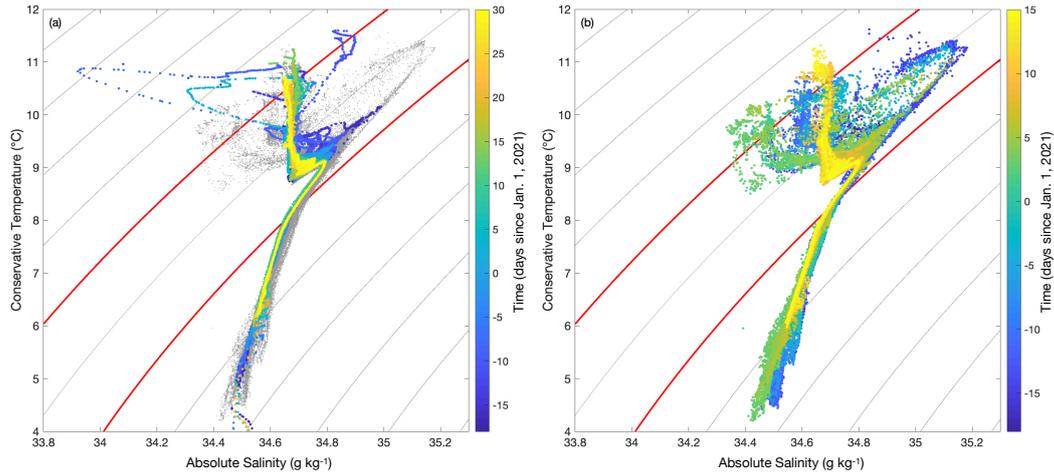


Figure 5. Distribution of conservative temperature Θ and absolute salinity S_A from the float and the glider. (a) Θ/S_A distribution from the float (colors) and glider (gray). (b) Θ/S_A distribution from the glider. Color in all panels indicates the time of the measurements given as days since January 1, 2021; note that the colorscales in the panels differ. The contours in each panel show isolines of constant potential density, σ_0 , with a contour interval of 0.2 kg m^{-3} ; the 26.6 and 27.0 kg m^{-3} contours are indicated by the red curves.

414 ues, with the coldest and freshest surface waters associated with the northern bound-
 415 ary of the SAF. The warmest and most saline waters are found a couple of hundred met-
 416 ers below the surface in the Θ - S_A compensated layer, associated with a southward flow-
 417 ing subsurface filaments of warm, salty subtropical waters; the strongest of these coincid-
 418 es with a filament between the northern anticyclone and western cyclone.

419 Variations in surface properties are correlated with the structure of the mesoscale
 420 flow field during the sampling period. In particular, mixed layer-averaged temperature
 421 and optical backscatter are elevated along the periphery of the coherent mesoscale fea-
 422 tures (Figure 6). These regions are also associated with the enhanced variability in MLD.
 423 The filament of warm and salty subtropical waters is collocated with an abrupt shoal-
 424 ing in the surface MLD as the float/glider enter region of high strain between the cy-
 425 clones. These cyclones have relatively cold interior waters, and therefore the mesoscale
 426 surface tracer gradients may allow for a cascade of tracer variance to smaller scales un-
 427 der mesoscale and submesoscale stirring. The mixed layer-integrated backscatter values
 428 have higher variability in the glider data. Peak backscatter values coincide with the tim-
 429 ing of the spring bloom (day -10 to -5, Figure 4) as well as during the submesoscale sam-
 430 pling period. The glider survey reveals $O(10)$ km-scale fronts in backscatter that are not
 431 apparent in the float data. The northern edge of the SAF and the large southern anti-
 432 cyclone have low backscatter values. Some of these variations might reflect the tempo-
 433 ral evolution of the bloom, but the small scale variability observed during the glider’s
 434 submesoscale survey likely reflects stirring of large scale surface variations by the mesoscale
 435 eddies (Lévy et al., 2018). These mesoscale eddies evolve over the month-long study pe-
 436 riod considered here, but the temporally averaged SSH contours shown in Figure 6 pro-
 437 vide a reliable map of eddy cores and straining regions at the eddy edges.

438 Interior hydrographic and biogeochemical distributions also differ between the strain-
 439 ing region and the latter part of the deployment when submesoscale variability is weaker
 440 (Figure 7). During the submesoscale survey, isopycals experience greater vertical displace-
 441 ments and there are large anomalies in spice and AOU tracers over scales of 10 km or

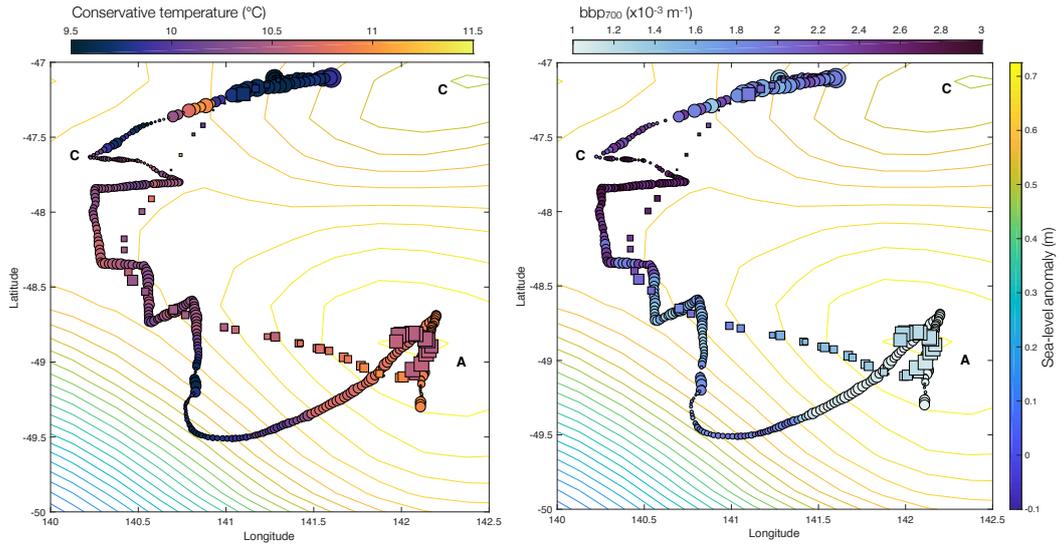


Figure 6. Surface mixed layer properties from the float and glider. (a) Mixed layer conservative temperature ($^{\circ}\text{C}$) overlaid on sea surface height contours (m), averaged between 15 December, 2020 and 15 January, 2021. The magnitude of the symbol is linearly proportional to the mixed layer depth. Labels A and C indicate the position of coherent anti-cyclones and cyclones. (b) As in panel (a) for for mixed layer averaged 700 nm optical backscatter (m^{-1}).

less (Figure 7,d-g), where AOU anomalies also remove a mean vertical AOU profile averaged over the submesoscale section (panels f,g). Notably, spice and AOU anomalies are largely uniform throughout the upper 400 m. The horizontal scales of variability suggest the potential for generating submesoscale instabilities, but the extension below the surface boundary layer implies the importance of mesoscale motions in generating these features. The second float-glider crossing (Figure 7e,g,i) indicates that within the core of the mesoscale eddies, there is less small-scale variability and gravitational sinking likely dominates the export flux in this region. Other float-glider crossings during the submesoscale survey show similar small-scale variability to Glider Section I in Figure 7; we return to the relationship between isopycnal variations in spice/AOU and backscatter distributions in section 5.1.

The vertical stratification, as measured by the glider, shows a complex pattern of layering as a deeper ($\sim 100\text{-m}$) mixed layer restratifies in late December and early January (Figure 8a). A persistent local maximum in buoyancy frequency, N^2 , is found at a depth of around 150 m. This depth is shallower than the 400-m wintertime mixed layer described by Rintoul and Trull (2001) (Figure 8b), but may represent mixed-layer deepening that occurs due to surface wind forcing. Early in the deployment, the upper ocean is weakly stratified and the MLD exceeds 100 m. Around yearday -12, the MLD shoals from greater than 100 m to < 20 m. This event is associated with a reduction in the magnitude of the latent heat flux that follows the passage of a strong synoptic wind event when the wind stress approached 0.8 N m^{-2} (Figure 3). The MLD shoaling is observed at the edge of the western coherent cyclonic eddy in a region of strong surface and subsurface lateral density gradients (Figure 8c). Over the next two weeks, MLD monotonically increased, despite a positive surface heat flux, until it reached a depth of roughly 70 m. Throughout the deployment, this shallow pycnocline is separated from the deeper permanent pycnocline by a layer of low stratification, a common feature in the Southern Ocean (du Plessis et al., 2019; Siegelman et al., 2020). During January, the MLD stabilized, but N^2 at the base of the mixed layer strengthened. During the brief period

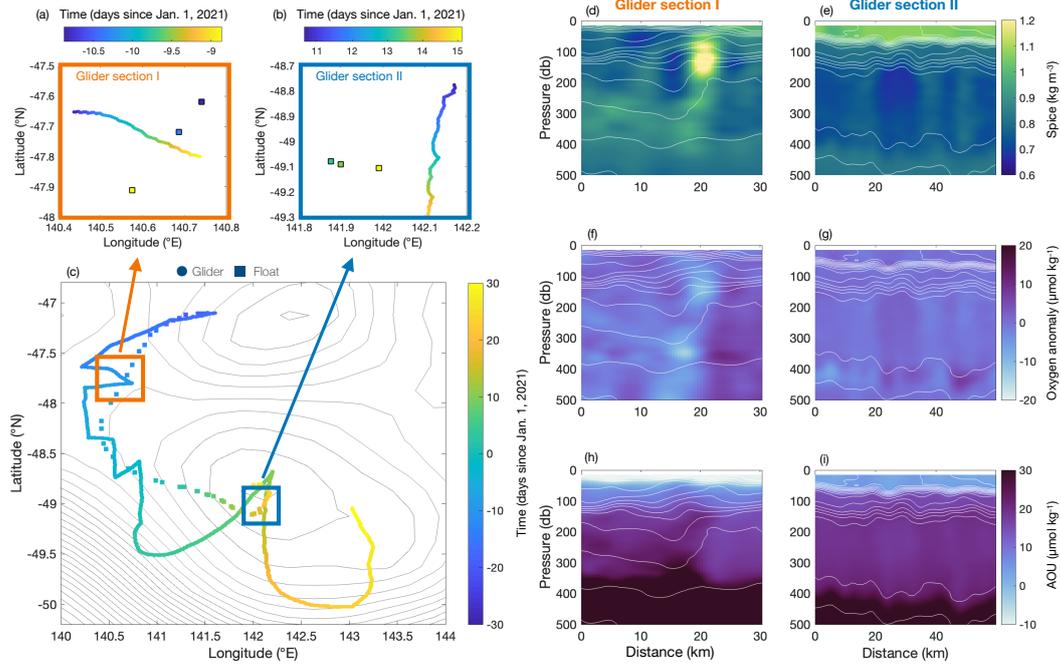


Figure 7. Submesoscale glider sections at float-glider crossovers. (a,b) Glider (circles) and float (squares) positions, colored by days since Jan. 1, 2021, at regions where they sampled in both spatial and temporal proximity. Glider section I (a) and section II (b) correspond to regions of dominant strain and vorticity, respectively (c); multiple crossovers occur in the higher strain region, as shown in panel (c). The contours in panel (c) are SSH anomaly as in Figure 1c. (d,e) Spice (kg m^{-3}), (f,g) dissolved oxygen anomaly ($\mu\text{mol kg}^{-3}$), and (h,i) apparent oxygen utilization (AOU, $\mu\text{mol kg}^{-3}$) for Glider section I (left panels) and II (right panels). The oxygen anomaly is determined by removing the mean value at each depth over the submesoscale section. The white contours in each panel indicate isopycnals with a 0.03 kg m^{-3} interval.

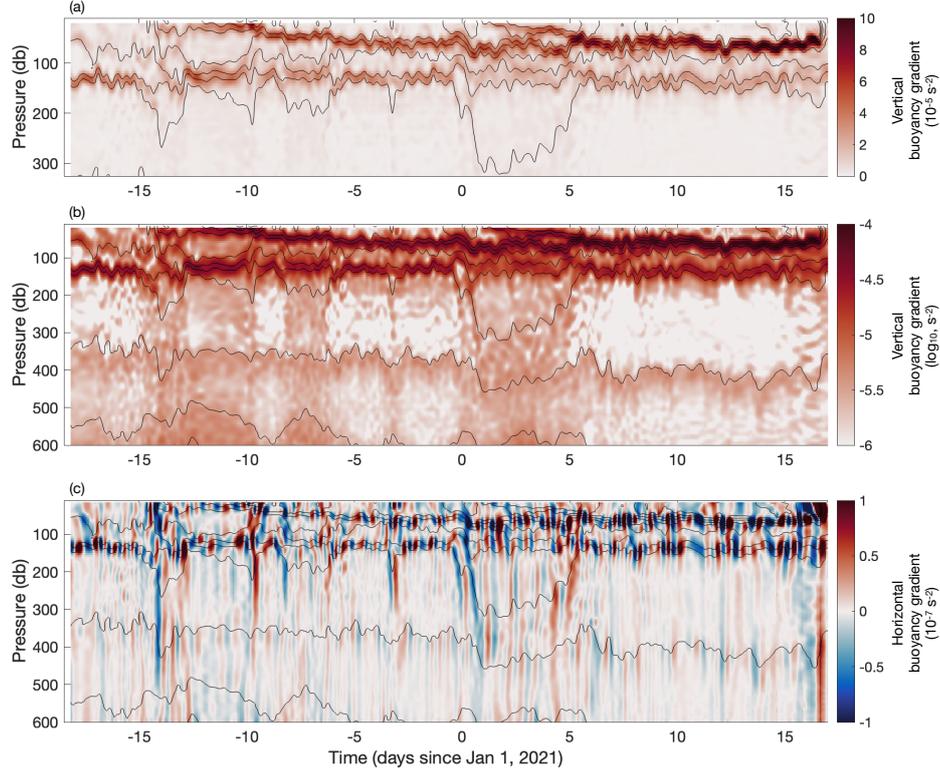


Figure 8. (a,b) Vertical stratification N^2 (defined in section 3) from the glider; note that panels (a) and (b) have different depth ranges (y-axis) and colorbars. (c) Lateral stratification M^2 (defined in section 3), along the trajectory of the glider. In all panels, the thin gray contours are isopycnals with a 0.05 kg m^{-3} contour interval.

470 when the glider sampled into the SAF, there are periods when three local maxima in N^2
 471 are found in a single vertical profile, potentially associated with the presence of subme-
 472 socale coherent vortices, or SCVs, subducting from the surface boundary layer. A layer
 473 of relatively weak stratification is also found between 150 and 400 m (Figure 8b), although
 474 the magnitude of the stratification in this layer is modulated by the ocean mesoscale,
 475 with regions of enhanced stratification associated with cyclonic vorticity.

476 Lateral density gradients, or M^2 , are dominated by oscillating positive and neg-
 477 ative anomalies localized at the seasonal and permanent pycnoclines — the signature of
 478 internal waves (Figure 8). These oscillations are weaker in the weakly stratified region
 479 between these layers. However, deeper and vertically-coherent signature in M^2 mark the
 480 edge of mesoscale eddies in the study region. These deep fronts are consistent with other
 481 Southern Ocean observations, *e.g.* from gliders and seals, (Viglione et al., 2018; Siegel-
 482 man et al., 2020), but notably occur in the SOLACE region, which is characterized by
 483 weaker EKE. Siegelman et al. (2020) argued that these deep lateral density gradients can
 484 support strong vertical velocities that extend to depths below the surface mixed layer.

485 4.2 Estimate of the mixed layer pump

486 The high-frequency sampling of the glider, roughly 10 vertical profiles per day, re-
 487 veals substantial variability in MLD, including changes of up to 50 m in one day dur-

488 ing the first week of the deployment. Daily fluctuations in MLD are damped following
 489 the onset of the seasonal pycnocline in late December, but the mixed layer continues to
 490 vary by many tens of meters on timescales of a few days to a week (Figure 9a). The float,
 491 sampling roughly once per day, captures a similar temporal evolution as the MLD (Fig-
 492 ure 10a); due to the quasi-Lagrangian nature of the float, these variations likely record
 493 the upper ocean’s response to atmospheric forcing (Figure 3). Variance in glider-derived
 494 MLD is elevated both when the mixed layer is deeper and in regions of mesoscale strain,
 495 conditions that are conducive to stronger submesoscale motions (Klein & Lapeyre, 2009).
 496 Both platforms record rapid shoaling events between yearday -17 and -15, although MLD
 497 deepens again to almost 100 m in the float time series. A longer estimate of the MLP
 498 export flux from the float has strong fluctuations in observed MLD in February and March
 499 2021, which is also associated with the float sampling along the periphery of a coherent
 500 cyclonic mesoscale eddy (not shown).

501 To estimate carbon export associated with the MLP, optical backscatter data from
 502 the float and the glider are converted to a POC concentration, and then vertically av-
 503 eraged and vertically integrated over the depth of the mixed layer for each profile (Fig-
 504 ure 9b and 10b). The mixed layer POC concentration gradually increases from day -20
 505 to day -10, peaking at $\sim 120 \text{ mg m}^{-3}$ during the spring bloom. From late December
 506 through January, the mixed layer POC concentration declines from 120 mg m^{-3} to 30 mg m^{-3} .
 507 The ML-integrated POC, on the other hand, exhibits finer temporal variations, reflect-
 508 ing fluctuations in MLD. The ML-integrated POC maximum occurs around day -17, ear-
 509 lier than the peak of the spring bloom, due to the deeper MLD. The differing behavior
 510 between mixed layer POC (increasing) and ML-integrated POC (decreasing) in mid-December
 511 suggests an important role for the mixed-layer pump on relatively short time scales.

512 Following Bol et al. (2018), daily and 5-day mean estimates of the MLP are cal-
 513 culated to produce a time series with a comparable temporal resolution to Bol’s year-
 514 long study from the North Atlantic (Figure 9c,d). Daily estimates of carbon export have
 515 a peak magnitude of over $2\text{-}4 \text{ g POC m}^{-2}$ and fluctuate between export (positive) and
 516 entrainment (negative) events. Averaging over a longer, 5-day, period, the daily export
 517 rates are smaller, with sustained periods of export/entrainment spanning one to two weeks,
 518 consistent with the MLD evolution recorded by the float. A sustained export event in
 519 mid-December, integrated over a 5-day period, reaches $2.25 \text{ g POC m}^{-2}$, although much
 520 of this export is counterbalanced by entrainment as the mixed layer deepens in late De-
 521 cember. Between yearday -20 and 15, the daily and 5-day glider estimates produce mean
 522 carbon export rates of -90.7 and $-39.2 \text{ mg POC m}^{-2} \text{ day}^{-1}$, respectively. Patterns of car-
 523 bon export and entrainment are broadly similar between the float and glider, which is
 524 expected since the glider was designed to sample water around the float (Figure 9, 10).
 525 Entrained and detrained POC concentrations were lower than the averaged mixed-layer
 526 POC concentrations, as was found by (Bol et al., 2018) (not shown), since POC concen-
 527 trations decrease with depth. The amplitude of MLP fluctuations is largest at the start
 528 of the deployment when the mixed layers are deepest and surface POC concentrations
 529 are high. The effectiveness of the MLP was influenced by the fact that peak surface POC
 530 concentrations occur after MLD shoals from its deepest values (Erickson & Thompson,
 531 2018). The timing of these two events differ by a week and would not be resolved by the
 532 typical 10-day Argo float sampling (Lacour et al., 2023).

533 4.3 Estimates of the eddy subduction pump

534 The generation of mixed layer density fronts has been identified as a potential mech-
 535 anism to enhance upper-ocean vertical velocities that can advect POC out of the sur-
 536 face boundary layer and into the ocean interior (Omand et al., 2015). The process most
 537 widely cited for generating these vertical motions is mixed layer baroclinic instability (Boccaletti
 538 et al., 2007), in which the formation of submesoscale eddies act to restratify, or “over-
 539 turn,” mesoscale density fronts. This overturning gives rise to both upward and down-

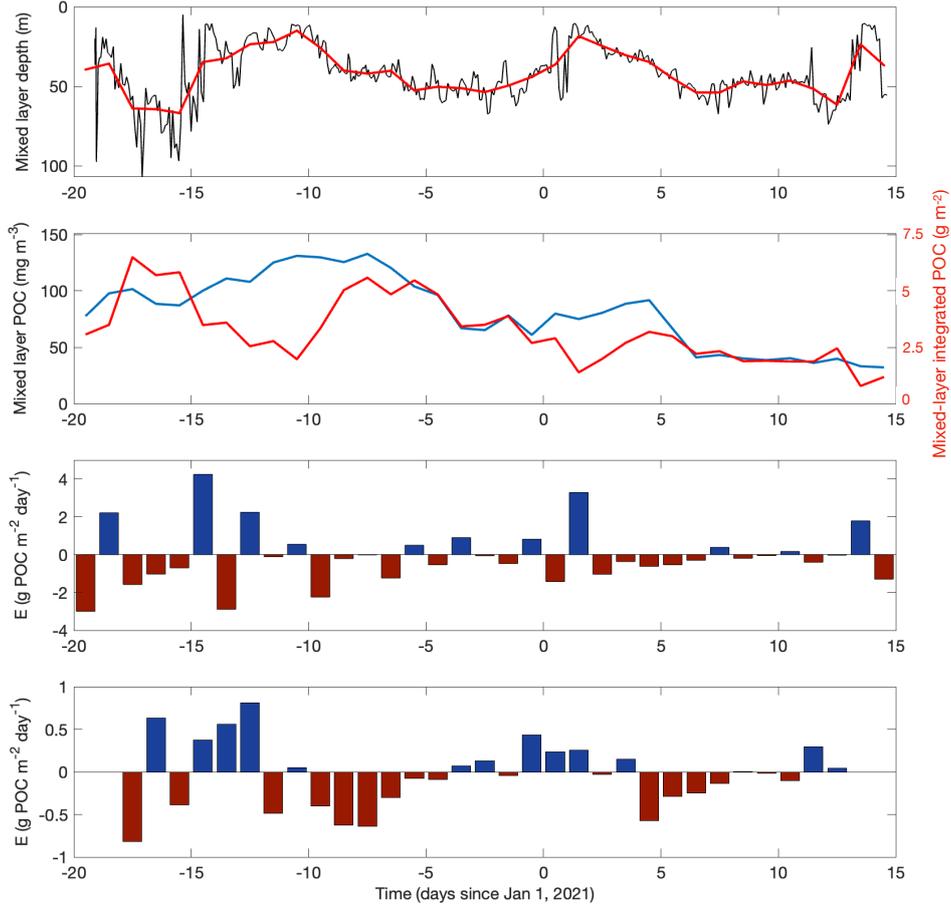


Figure 9. High-frequency mixed layer pump estimate. (a) Time series of mixed layer depth from each glider profile (black) and the glider daily-mean value (red). (b) Time series of mixed layer POC concentration (blue curve, mg m^{-3}) and POC concentration integrated over the depth of the mixed layer (red curve, g m^{-2}). (c) Daily-averaged estimate of the mixed layer pump E ($\text{g POC m}^{-2} \text{day}^{-1}$) as defined in equation (5) in section 3. (d) Five-day-averaged estimate of the mixed layer pump E ($\text{g POC m}^{-2} \text{day}^{-1}$) as defined in equation (5). In panels (c) and (d), positive and negative values are associated with POC export from and entrainment into the mixed layer, respectively.

ward motions, on the lighter and denser sides of the fronts respectively (Fox-Kemper et al., 2008; Taylor & Thompson, 2023), but supports a net downward flux of POC, assuming POC concentration decreases with depth. The scaling for the eddy subduction carbon flux in equation (3) does not explicitly resolve upwelling and downwelling components, although both observations and models provide evidence that downward POC and chlorophyll fluxes are enhanced on the dense side of fronts (Freilich & Mahadevan, 2021). We apply the scaling here with a focus on spatial variations in the magnitude of the ESP.

During the month-long deployment, variations in upper ocean stratification, both N^2 and M^2 , make the dominant contribution to the temporal variability in the estimated F_{ESP} (Figure 11). As discussed above, the mixed layer undergoes a rapid shoaling event around day -17, but subsequent MLD variability is more muted (Figure 11a). Mixed layer baroclinic instability is an adiabatic process, with particles assumed to be advected along isopycnals. Thus, an increase in the tilt of density surfaces, where the isopycnal slope

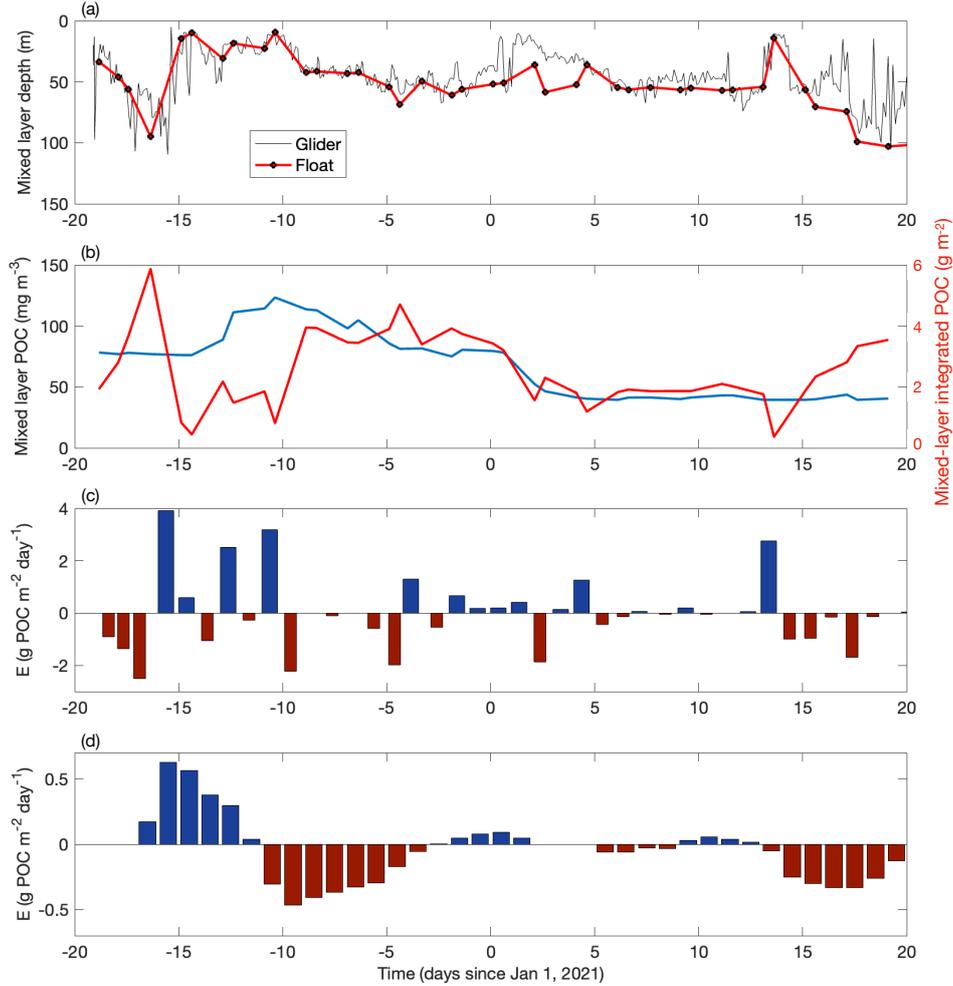


Figure 10. As in Figure 9, but for the float.

553 is given by the ratio M^2/N^2 , are more conducive to export. Lateral buoyancy gradients
 554 are elevated earlier in the deployment, even during times, *e.g.* days -15 to -10, when the
 555 MLD is shallow (Figure 11c). The period of enhanced M^2 corresponds with times of peak
 556 N^2 during days -15 through -10 (Figure 3a), which may be a signature of effective re-
 557 stratification by submesoscale eddies. The strong stratification at the base of the mixed
 558 layer differs substantially from typical North Atlantic wintertime and early springtime
 559 conditions where submesoscale fluxes have been suggested as being important for carbon
 560 export (Omand et al., 2015; Erickson & Thompson, 2018).

561 The mixed layer POC concentration, estimated from the optical backscatter mea-
 562 surements as described in section 3, are elevated throughout most of December, tran-
 563 sitioning to moderate values around December 28. Around January 7, POC concentra-
 564 tion drops rapidly to small values, consistent with the low values of surface backscatter
 565 measured by the float in the southern anticyclone (Figure 11d). Physical properties of
 566 the upper ocean are combined to estimate an eddy streamfunction that peaks early in
 567 the deployment when mixed layers are deepest. However, the export flux depends on both
 568 the overturning strength and the isopycnal slope. Thus, even as the eddy streamfunc-
 569 tion weakens in late December, F_{ESP} remains intermittently $> 500 \text{ mg m}^{-2} \text{ day}^{-1}$, and
 570 peaks at almost $700 \text{ mg m}^{-2} \text{ day}^{-1}$, around day -10; over the next few days, F_{ESP} re-

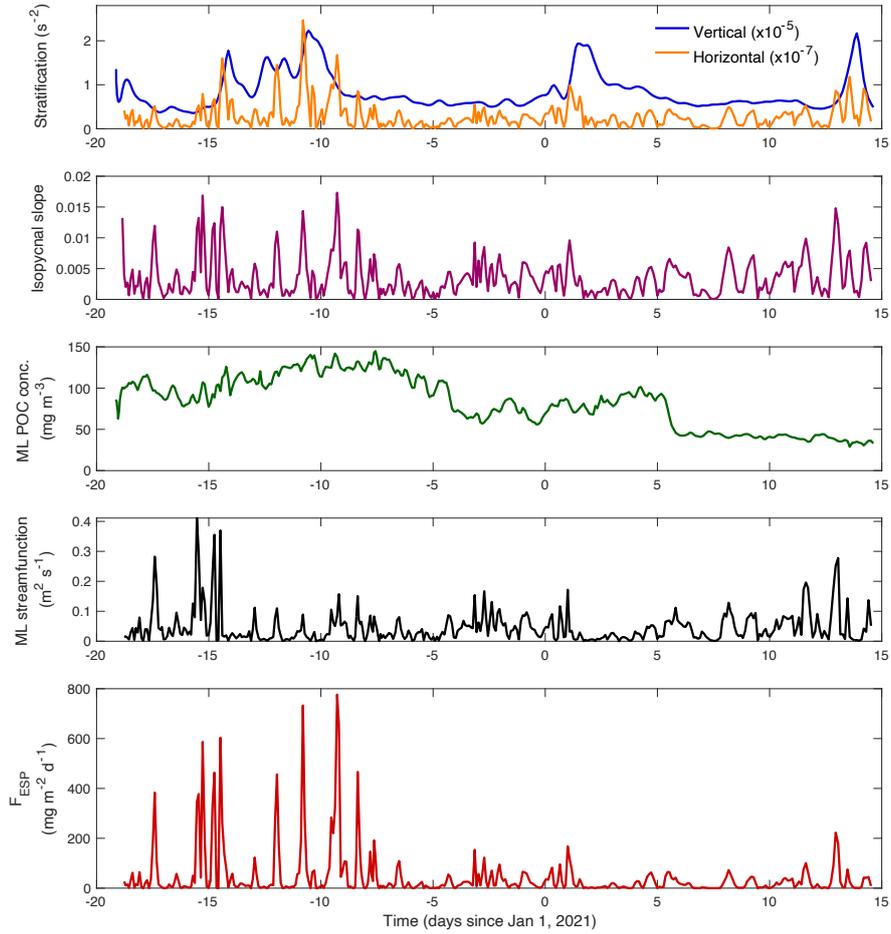


Figure 11. Estimation of the mixed layer pump along the trajectory of the glider. Time series of (a) vertical stratification, N^2 (s^{-2}), and lateral stratification M^2 (s^{-2}), integrated over the depth of the mixed layer (ML); (b) isopycnal slope, N^2/M^2 ; (c) POC concentration within the mixed layer (mg m^{-3}); (d) parameterized mixed layer eddy streamfunction (see discussion in section 3); and (e) time series of POC export following equation (3).

571 mains elevated due to a weakening N^2 . This 5-day period (yearday -12 to -7) corresponds
 572 to the first part of the submesoscale survey and increased variability in surface proper-
 573 ties consistent with the presence of submesoscale anomalies or stirring by submesoscale
 574 eddies. This somewhat intricate dependence of both the magnitude and phasing of mul-
 575 tiple upper ocean properties indicates that contribution of the ESP to carbon export in
 576 any given season can be sensitive to small shifts in the surface forcing and its impact on
 577 MLD, vertical stratification, and the strength of frontogenetic and straining fields from
 578 nearby mesoscale eddies (Erickson & Thompson, 2018). Over the period, day -20 to 15,
 579 the mean value of F_{ESP} is $153 \text{ mg m}^{-2} \text{ day}^{-1}$, and thus likely makes a stronger contri-
 580 bution to export than the MLP during the SOLACE field program.

581 5 Discussion

582 5.1 Isopycnal variability and the eddy subduction pump

583 Dynamical processes associated with coherent eddies and fronts can strongly in-
 584 fluence biogeochemical tracer distributions and carbon export. Studies have often focused
 585 specifically on mesoscale, *e.g.*, (McGillicuddy Jr. et al., 1998; Gaube et al., 2014), or sub-
 586 mesoscale, *e.g.* (Lévy et al., 2012; Mahadevan, 2016; Omand et al., 2015; Estapa et al.,
 587 2015; Archer et al., 2020), processes. Yet, there is growing evidence that the magnitude
 588 and variability of export fluxes depend on processes that span these scales. A rising paradigm
 589 is that mesoscale strain and stirring establish strong, near-surface lateral density gra-
 590 dients via frontogenesis (Balwada et al., 2018; Su et al., 2020) and enhance vertical ve-
 591 locities due to surface convergence (Freilich & Mahadevan, 2021). These processes give
 592 rise to ageostrophic vertical velocities that are a key component of transport across the
 593 base of the mixed layer (McWilliams, 2016; Dove et al., 2021; Taylor & Thompson, 2023).
 594 Further subduction of surface water properties in the interior is dominated by mesoscale
 595 stirring (Freilich & Mahadevan, 2021), although large submesoscale vertical velocities have
 596 been inferred at depth in energetic regions of the ocean (Siegelman et al., 2020; Yu et
 597 al., 2019). Here, we examine the potential for small-scale variations in the ESP in a re-
 598 gion of the Southern Ocean with relatively low EKE, following the paradigm outlined
 599 above.

600 Eddy subduction has been linked to enhanced isopycnal tracer variance of spice,
 601 an (essentially) passive tracer (Dove et al., 2021), as well as deep anomalies of low AOU
 602 waters (Llort et al., 2018). Near the SOTS region, AOU has a non-monotonic distribu-
 603 tion when plotted against density due to interleaving between high AOU waters located
 604 to the north and relatively low AOU waters associated with AAIW (Figure 12a). This
 605 interleaving spans a vertical depth of ~ 300 m, but is compact in density space, occur-
 606 ring in the range $26.8 < \sigma_0 < 26.85 \text{ kg m}^{-3}$. This density layer hosts lateral gradi-
 607 ents in AOU that are comparable to the vertical gradient in AOU (Figure 12b), vary-
 608 ing by $\sim 40 \mu\text{mol kg}^{-1}$ over a distance of 10 km. These submesoscale anomalies in AOU
 609 are noticeably enhanced during the glider’s submesoscale survey, coincident with the strongest
 610 straining region sampled by the float and glider (Figure 7). Here, anomalies are mostly
 611 vertically coherent throughout the upper 1000 m, which suggests a key role for mesoscale
 612 stirring. The anomalies are notably two to three times smaller than isopycnal AOU anom-
 613 alies observed in strong stirring regions in the lee of topographic features in the ACC (Llort
 614 et al., 2018; Dove et al., 2021). The shallow sampling pattern of the float only marginally
 615 sampled the strong oxycline associated with the $\sigma_0 = 26.8 \text{ kg m}^{-3}$ isopycnal surface
 616 (not shown).

617 The distribution of spice and optical backscatter offer additional insight into the
 618 interplay between mesoscale and submesoscale features. Spice variance is dominated by
 619 relatively small-scale structures associated with straining regions found along the periph-
 620 ery of mesoscale eddies and fronts (Figure 13a,c). These anomalies are strongest above
 621 the the $\sigma_0 = 26.8 \text{ kg m}^{-3}$ isopycnal surface, but are largely found below the base of the

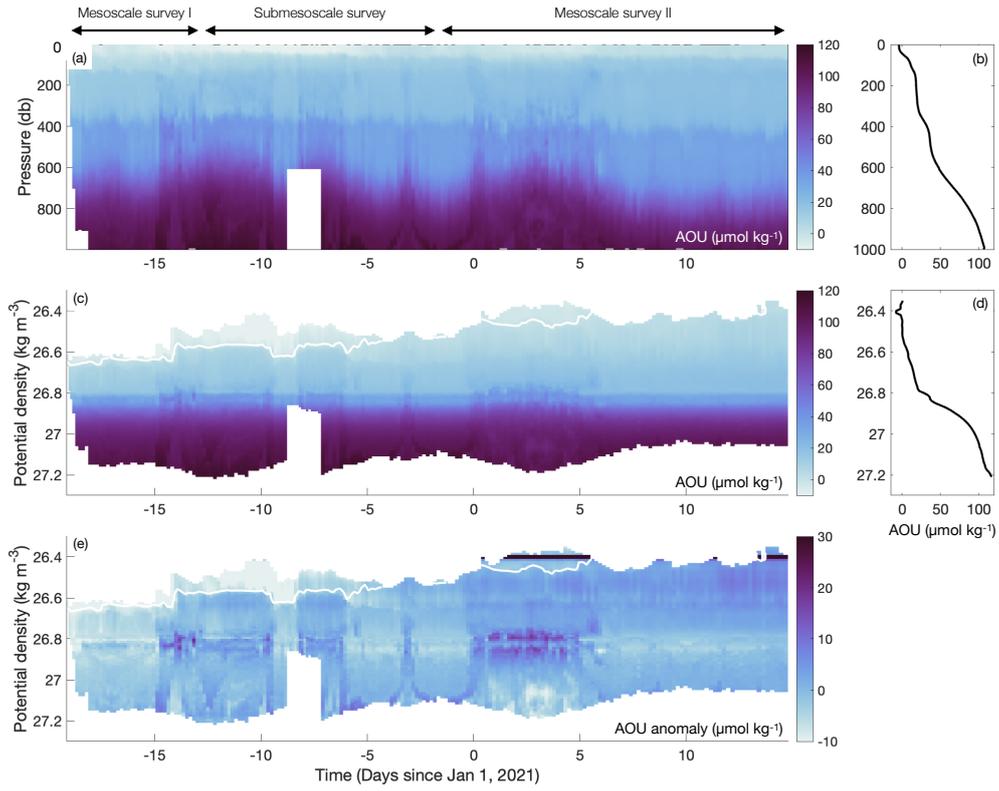


Figure 12. Apparent oxygen utilization (AOU, $\mu\text{mol kg}^{-3}$) variability along the trajectory of the glider. (a) Depth-time section of AOU, and (b) the time-mean distribution of AOU with depth covering the period shown in panel (a). (c) Density-time section of AOU, and (d) the time-mean distribution of AOU with potential density during the period shown in panel (c). (e) Density-time section of AOU anomalies, after removing the mean AOU profile shown in panel (d). The white contours in panels (c) and (e) correspond to AOU = 0 $\mu\text{mol kg}^{-3}$.

622 mixed layer, consistent with these features arising from mesoscale stirring, rather than
 623 submesoscale processes. Spice isolates stirring dynamics in this region, while optical backscat-
 624 ter integrates the effects of both advection and gravitational sinking, explaining differ-
 625 ences in their isopycnal distributions. In depth space, backscatter has a sharp vertical
 626 gradient at the base of the mixed layer (Figure 13b), with variations across both the sea-
 627 sonal pycnocline and the permanent pycnocline. In density space, there is a strong ver-
 628 tical gradient in optical backscatter at the $\sigma_0 = 26.8 \text{ kg m}^{-3}$ isopycnal surface, and op-
 629 tical backscatter variability is particularly pronounced in the density range $26.6 < \sigma_0 <$
 630 26.8 kg m^{-3} (Figure 13d). Elevated backscatter values occur in this density range be-
 631 tween days -9 to -8 and days -6 to -3 , for example. When considering spice and backscat-
 632 ter distributions together, these backscatter anomalies are bookended by the high spice
 633 values.

634 The optical backscatter distribution highlights the coupling between mesoscale and
 635 submesoscale dynamics on carbon export. First, periods when backscatter is elevated
 636 in the $26.6 < \sigma_0 < 26.8 \text{ kg m}^{-3}$ density range coincide with the displacement of isopy-
 637 cnals associated with coherent mesoscale cyclonic vorticity and the ACC's SAF (Figure
 638 13a). The shoaling of the density surfaces allows this density class to be preferentially
 639 "imprinted" with sinking particles that can be efficiently stirred along density surfaces
 640 due to mesoscale stirring. As the particulate matter is advected along isopycnals, slop-
 641 ing density surfaces at the periphery of eddies provide pathways for enhanced carbon ex-
 642 port that is not linked to submesoscale motions, *e.g.*, $Ro = O(1)$ dynamics, but occurs
 643 at scales smaller than the coherent mesoscale eddy. A striking feature of the backscat-
 644 ter distribution is that a subsurface anomaly develops along the $\sigma_0 \approx 27.5 \text{ kg m}^{-3}$ sur-
 645 face, following these isopycnal shoaling events (Figure 13d-f). This injection of partic-
 646 ulate matter along these density surfaces, in some cases, crosses into depth strata where
 647 there is a marked attenuation in particle flux in the BGP (Grabowski et al., 2019)

648 To estimate a subduction timescale for this process, $\tau_{\text{sub.}}$, we apply a simple scal-
 649 ing analysis, $\tau_{\text{sub.}} \sim (ds_\rho^{-1})^2 K^{-1}$, where d is the subduction depth, K is an eddy dif-
 650 fusivity, and $s_\rho \equiv M^2/N^2$ is the isopycnal slope. Taking $d = 100 \text{ m}$, and applying typ-
 651 ical diffusivity values for the Southern Ocean, $K \approx 100\text{-}1000 \text{ m}^2 \text{ s}^{-1}$, and observed val-
 652 ues $s_\rho \approx 0.01$ (Figure 11), subduction of POC anomalies 100 m vertically through the
 653 water column can occur over 1 to 10 days.

654 This mechanism, summarized in the schematic in Figure 14c, is distinct from the
 655 rapid subduction of particulate matter by enhanced submesoscale advection linked to
 656 the ESP (Ruiz et al., 2009; Llort et al., 2018). Subduction via the ESP is more consis-
 657 tent with the backscatter distributions found around day -13 and discussed in more de-
 658 tail in the next section. We note that both gravitational sinking and shallower subme-
 659 soscale motions may be key for seeding certain density classes with high levels of par-
 660 ticulate matter that can be redistributed vertically and horizontally by interior mesoscale
 661 stirring.

662 5.2 Separating contributions from mixed layer and eddy subduction pumps

663 The two PPIPs considered in this study, the MLP and ESP, each have a strong de-
 664 pendence on MLD and upper ocean stratification. For the MLP, a reduction in MLD,
 665 originally envisioned as due to surface forcing (Gardner et al., 1995), results in a pas-
 666 sive removal of POC from the surface boundary layer and, through subsequent sinking,
 667 an escape from future re-entrainment when the mixed layer deepens again. In contrast,
 668 the ESP mechanism involves the generation of surface density gradients that support strong
 669 vertical velocities and an active advection of POC from the surface boundary layer. The
 670 signature of export by the ESP is difficult to distinguish from sparse, often one- or two-
 671 dimensional, observations, especially when subsurface POC anomalies are displaced lat-
 672 erally from the location where they exited the mixed layer. Additionally, mixed layer baro-

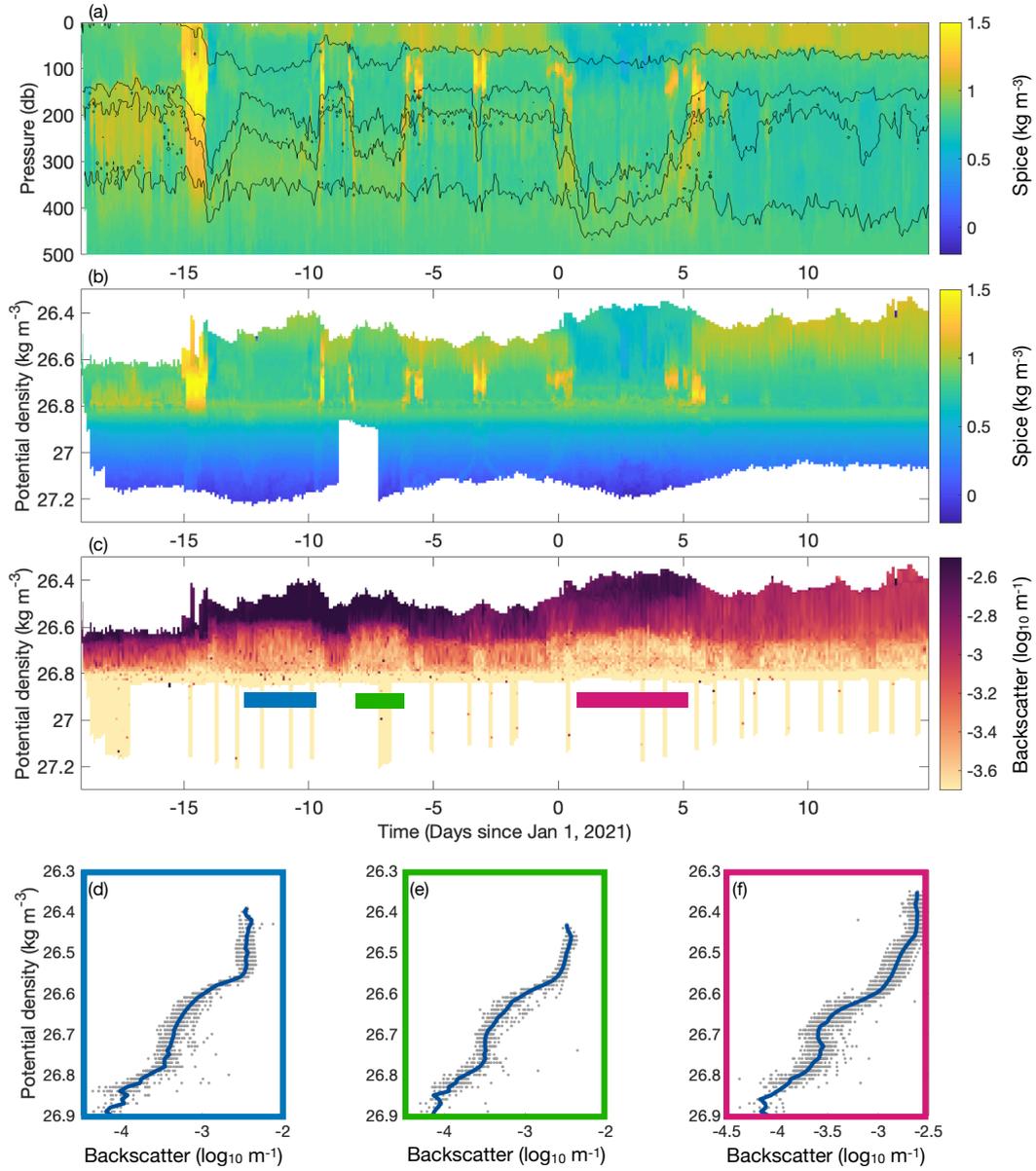


Figure 13. (a) Spice (color) and potential density, σ_0 (contours) along the glider trajectory, both with units kg m^{-3} ; the σ_0 contours are 26.6, 26.76, 26.78, and 26.8 kg m^{-3} . (b) Spice and (c) optical backscatter (bbp, m^{-1}) mapped on to potential density surfaces with a 0.01 kg m^{-3} interval. Lower panels: time-averaged bbp during the year day intervals, -13 to -10 (panel d, blue), -8 to -6 (panel e, green), and 0 to 5 (panel f, magenta).

673 clinic instability leads to a shoaling and restratification of the mixed layer, which can
674 make attribution of carbon export from an individual PPIP challenging.

675 The first half of the SOLACE deployment highlights two distinct scenarios in which
676 MLP and ESP pumps interact. The export flux event centered around day -15 is asso-
677 ciated with (i) a rapid shoaling of the mixed layer, (ii) peak values of the eddy overturn-
678 ing streamfunction (Figure 11) and (iii) a deep-reaching density front in the high strain
679 region between the two cyclones (Figure 1). Furthermore, in mid-December 2020 this
680 region was characterized by a diverse phytoplankton assemblage, from small pico-eukaryotes
681 to large diatoms, with a high photosynthetic competence (F_v/F_m), and hence cells likely
682 had negligible sinking rates (E. Shadwick, personal communication). This suggests that
683 injection to depth via physically-mediated injection pumps was the dominant export path-
684 way.

685 As discussed in section 5.1, day -15 marks the initiation of enhanced interior tracer
686 gradients (spice and apparent oxygen utilization), but most of the shoaling of the mixed
687 layer happens one or two days earlier. The mixed layer shoaling follows a reduction in
688 the surface wind stress, a reduction in the magnitude of the latent heat flux, and an in-
689 crease in the total surface heat flux (Figure 3). This is consistent with carbon export in
690 response to 1D surface forcing. Thus, during this period, any contribution to POC ex-
691 port from the ESP would have occurred due to the surface-forcing induced mixed-layer
692 shoaling, or the MLP. In this scenario, it is not appropriate to sum the two PPIP con-
693 tributions. We propose that the PPIP contribution is likely to be better represented by
694 the maximum of carbon flux estimated from the MLP and ESP parameterizations (Fig-
695 ure 14a). For this study, the estimated contributions from the MLP and ESP are of com-
696 parable magnitude during this short period.

697 The ESP has two strong peaks, one occurring between days -17 to -15, associated
698 with deep mixed layers, and the second between days -12 to -8, associated with relatively
699 shallow mixed layers but enhanced lateral density gradients and a weakened vertical strat-
700 ification N^2 . In the latter period, the ESP and MLP work in opposition, with $F_{\text{ESP}} >$
701 0 and $F_{\text{MLP}} < 0$. Frontogenesis occurring in this straining region provides conditions
702 that are conducive to strengthening F_{ESP} , even as the the mixed layer slowly deepens,
703 entraining potentially subducted POC. The two PPIPs cancel to leading order, although
704 averaging over this four day period gives a slight tendency for export (Figure 14b). The
705 mixed layer deepening is likely a broad response to wind-driven mixing since increasing
706 MLD is recorded in both the float and glider data, whereas subduction via the ESP is
707 more spatially localized.

708 A limitation of the Omand et al. (2015) scaling is that the vertical structure of the
709 ESP is not explicitly addressed. If the vertical advection is sufficiently deep that par-
710 ticulate matter escapes subsequent deepening of the mixed layer, then this export would
711 not be offset by the MLP (Figure 14b). These interactions highlight a major challenge
712 with defining appropriate export horizons for export flux calculations if both PPIPs are
713 active: the MLP is best assessed in depth, relative to the depth of the mixed layer base,
714 whereas the ESP is easier to diagnose in density space.

715 6 Conclusions

716 The SOLACE data set, as a whole, provides a unique opportunity to estimate the
717 combined contributions from multiple particle injection pumps to carbon export in the
718 Subantarctic Zone of the Southern Ocean. This study (i) provides contextual hydrographic
719 and biogeochemical observations from the region due to the broad sampling of the float-
720 glider observing pair, and (ii) conducts a first attempt to isolate contributions from two
721 PPIPs, the MLP and ESP. While the SOTS region is a relatively low EKE region of the
722 Southern Ocean, there is compelling evidence that physical processes spanning meso- and

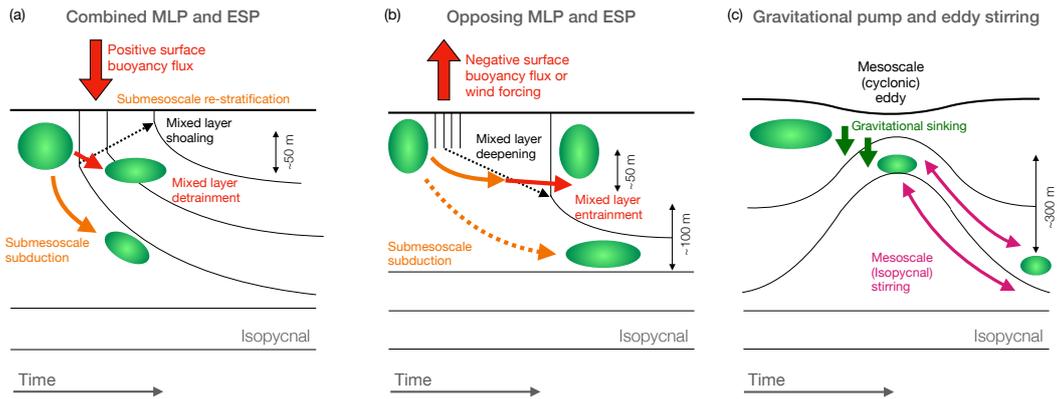


Figure 14. Examples of combined particle injection pumps. (a) Positive surface buoyancy forcing in the presence of submesoscale surface fronts can cause carbon export through both eddy subduction and detrainment (mixed layer shoaling); these processes may be difficult to separate unless subduction injects carbon to deeper depths and density classes. (b) Eddy subduction may occur during periods of entrainment (mixed layer deepening); the ESP and MLP would cancel unless subduction injects carbon to depths where it is not re-entrained. (c) Mesoscale eddies can shoal density surfaces, allowing gravitational sinking to intermittently imprint particles on the deeper density classes; subsequent along-isopycnal stirring, which is enhanced at the eddy periphery, can efficiently carry these particles to depth.

723 submesoscales make a significant contribution to the export of carbon following the
 724 onset of the spring bloom. The timing of mixed layer restratification coincided with the
 725 float/glider occupying a region of mesoscale strain and associated enhanced lateral den-
 726 sity gradients, such that both the MLP and ESP were likely active and contributing to
 727 the export of the same carbon stock. This highlights some of the potential complications
 728 with simply summing carbon fluxes estimated from PPIPs in isolation.

729 While the primary focus of this study was on PPIPs, the interior shaping of the
 730 density surfaces by mesoscale eddies are also shown to influence the vertical transport
 731 of POC. In regions of cyclonic vorticity, density classes that are typically many hundreds
 732 of meters below the base of the mixed layer shoal by $O(100\text{ m})$ or more and are exposed
 733 to higher POC concentrations as a result of gravitational sinking. Sampling with a single
 734 glider and quasi-Lagrangian float does not permit direct tracking of these POC anom-
 735 alies, but assuming typical mesoscale isopycnal stirring diffusivities, $100\text{-}1000\text{ m}^2\text{ s}^{-1}$ and
 736 an isopycnal slope of 0.01, vertical displacements of $\sim 100\text{ m}$ may occur on the order of
 737 days. This coupling between gravitational and advective fluxes will be strongest in sub-
 738 mesoscale regions on the periphery of eddies where strain and isopycnal tilt is enhanced.

739 In addition to the challenge of distinguishing carbon flux contributions from various
 740 PPIPs using sparse observations, a theoretical framework for determining global and
 741 annual contributions from these PPIPs is daunting. Selecting an appropriate export hor-
 742 izon, or the depth below organic carbon is considered exported (Palevsky & Doney, 2018),
 743 to estimate an export flux may be sensitive to not only local mixed layer depth, but also
 744 the stratification at the base of the mixed layer, the range of densities that outcrop in
 745 the mixed layer, and the interior density structure. Future efforts should, through a com-
 746 bination of observational analysis and high-resolution process-based modeling, focus on
 747 improving mechanistic understanding of how the vertical structure of physical and bio-
 748 geochemical properties relate to export flux magnitudes and the relevant carbon pumps.

749 7 Open Research

750 Surface flux and forcing data is available through the Australian Ocean Data Net-
 751 work at <https://portal.aodn.org.au/>. Sea level anomaly and surface velocity prod-
 752 ucts were produced and distributed by the Copernicus Marine 360 and Environment Mon-
 753 itoring Service and are available at [https://data.marine.copernicus.eu/product/
 754 SEALEVEL_GLO_PHY_L4_MY_008_047/description](https://data.marine.copernicus.eu/product/SEALEVEL_GLO_PHY_L4_MY_008_047/description). Finite-size Lyapunov Exponents were
 755 produced and distributed by AVISO+ and are available at [https://www.aviso.altimetry
 756 .fr/en/data/products/value-added-products/fsle-finite-size-lyapunov-exponents
 757 .html](https://www.aviso.altimetry.fr/en/data/products/value-added-products/fsle-finite-size-lyapunov-exponents.html)

758 Glider data presented in this work has been uploaded to the NOAA National Cen-
 759 ters for Environmental Information (NCEI) database and is accessible at [https://www
 760 .ncei.noaa.gov/archive/accession/0276999](https://www.ncei.noaa.gov/archive/accession/0276999).

761 Float data can be downloaded from the Argo Global Data Assembly Center ([ftp://
 762 ftp.ifremer.fr/ifremer/argo/](ftp://ftp.ifremer.fr/ifremer/argo/)). These data were collected and made freely available
 763 by the International Argo Program and the national programs that contribute to it: ([http://
 764 www.argo.ucsd.edu](http://www.argo.ucsd.edu), <https://www.ocean-ops.org>). The Argo Program is part of the
 765 Global Ocean Observing System.

766 Acknowledgments

767 We are grateful for the efforts of the full SOLACE field team as well as the captain and
 768 crew of the R/V Investigator. We extend particular thanks to Tyler Rohr who took on
 769 extra glider testing, deployment and troubleshooting responsibilities during the cruise
 770 due to COVID travel restrictions. The authors acknowledge helpful suggestions and con-
 771 versations from Elizabeth, Shadwick, Tom Trull, Mara Freilich, and Alice Della Penna.
 772 AFT, LAD and EF were supported by the Resnick Sustainability Institute and the Ginkgo
 773 Foundation; LAD received additional support from an NSF graduate research fellowship.
 774 EF's contributions were supported by a Summer Undergraduate Research Fellowship at
 775 Caltech. LL was supported by a European Union's Horizon 2020 Marie Skłodowska-Curie
 776 grant (no. 892653).

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