



HAL
open science

Submesoscales Enhance Storm-Driven Vertical Mixing of Nutrients: Insights From a Biogeochemical Large Eddy Simulation

D. B. Whitt, Marina Lévy, John R. Taylor

► **To cite this version:**

D. B. Whitt, Marina Lévy, John R. Taylor. Submesoscales Enhance Storm-Driven Vertical Mixing of Nutrients: Insights From a Biogeochemical Large Eddy Simulation. *Journal of Geophysical Research. Oceans*, 2019, 124 (11), pp.8140-8165. 10.1029/2019JC015370 . hal-02422848

HAL Id: hal-02422848

<https://hal.science/hal-02422848>

Submitted on 8 Dec 2020

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 Submesoscales enhance storm-driven vertical mixing
2 of nutrients: insights from a biogeochemical large
3 eddy simulation

D. B. Whitt¹, M. Lévy², and J. R. Taylor³

Corresponding author: D. B. Whitt, National Center for Atmospheric Research, 1850 Table Mesa Dr, Boulder, CO, USA. (dwhitt@ucar.edu)

¹National Center for Atmospheric Research, Boulder, CO, USA.

²Sorbonne Université
CNRS/IRD/MNHN, LOCEAN-IPSL, Paris,
France.

³Department of Applied Mathematics and Theoretical Physics, Centre for Mathematical Sciences, University of Cambridge, Cambridge, United Kingdom.

4 **Abstract.** Storms deepen the mixed layer, entrain nutrients from the py-
5 cnocline, and fuel phytoplankton blooms in mid-latitude oceans. However,
6 the effects of oceanic submesoscale (0.1-10 km horizontal scale) physical het-
7 erogeneity on the physical-biogeochemical response to a storm are not well
8 understood. Here, we explore these effects numerically in a Biogeochemical
9 Large Eddy Simulation (BLES), where a four-component biogeochemical model
10 is coupled with a physical model that resolves some submesoscales and some
11 smaller turbulent scales (2 km to 2 m) in an idealized storm forcing scenario.
12 Results are obtained via comparisons to BLES in smaller domains that do
13 not resolve submesoscales and to one-dimensional column simulations with
14 the same biogeochemical model, initial conditions, and boundary conditions
15 but parameterized turbulence and submesoscales. These comparisons show
16 different behaviors during and shortly after the storm. During the storm, re-
17 solved submesoscales double the vertical nutrient flux. The vertical diffusiv-
18 ity is increased by a factor of 10 near the mixed layer base, and the mixing-
19 induced increase in potential energy is double. Resolved submesoscales also
20 enhance horizontal nutrient and phytoplankton variance by a factor of 10.
21 After the storm, resolved submesoscales maintain higher nutrient and phy-
22 toplankton variance within the mixed layer. However, submesoscales reduce
23 net vertical nutrient fluxes by 50% and nearly shut off the turbulent diffu-
24 sivity. Over the whole scenario, resolved submesoscales double storm-driven
25 biological production. Current parameterizations of submesoscales and tur-

26 bulence fail to capture both the enhanced nutrient flux during the storm and
27 the enhanced biological production.

1. Introduction

28 In the mid-latitudes, strong winds associated with storms intermittently enhance upper-
29 ocean turbulence and hence vertical mixing of nutrients and phytoplankton [e.g., Marra
30 et al., 1990; Rumyantseva et al., 2015; Carranza & Gille, 2015; Nicholson et al., 2016].
31 These storm events can cause significant intra-seasonal variability of upper-ocean phyto-
32 plankton concentrations and therefore significant divergence from the mean seasonal cycle
33 [e.g., Waniek, 2003; Thomalla et al., 2011]. These variations in phytoplankton phenology
34 impact the higher trophic levels of marine ecosystems [e.g., Cushing, 1990; Platt et al.,
35 2003].

36 An unresolved question is to what degree quasi two-dimensional ocean submesoscales
37 (defined here descriptively to refer to features with horizontal length scales from 0.1 to
38 10 km, and small aspect ratio) interact with smaller-scale wind-driven three-dimensional
39 upper-ocean turbulence (< 100 m) to modify the physical-biogeochemical response to a
40 storm. This study addresses this question in the oceanic context of a fairly typical autumn
41 storm that was observed in the mid-latitude North Atlantic at 48.78°N , 16.38°W during
42 the OSMOSIS project [e.g., Rumyantseva et al., 2015; Painter et al., 2016] by showing how
43 submesoscales impact net (submesoscale + turbulent) vertical nutrient flux and resulting
44 biological production during and shortly after the passage of the storm in an ocean model.

45 In the absence of storms, prior work has shown that submesoscales can both enhance
46 vertical tracer fluxes [e.g., Mahadevan & Archer, 2000; Lévy et al., 2001; Capet et al.,
47 2008b] or suppress vertical fluxes by restratifying the ocean mixed layer or reducing the
48 depth of the surface mixing layer in ocean models [e.g., Boccaletti et al., 2007; Fox-Kemper

49 et al., 2008; Mahadevan et al., 2012; Bachman et al., 2017b]. Enhanced vertical fluxes, for
50 example, can induce a net increase in the nutrient delivery to the well-lit euphotic layer
51 and therefore a net increase in phytoplankton growth in oligotrophic oceans [Mahadevan
52 & Archer, 2000; Lévy et al., 2001].

53 During storms, winds can increase the vertical nutrient flux (thereby enhancing net
54 phytoplankton growth in oligotrophic oceans), both by enhancing submesoscale vertical
55 velocities [e.g., Mahadevan & Tandon, 2006; Capet et al., 2008a; Thomas et al., 2008;
56 Mahadevan et al., 2008; Lévy et al., 2009; Brannigan, 2016] and via entrainment/mixing
57 in submesoscale fronts [e.g., Lévy et al., 2009; Whitt et al., 2017b]. In addition, winds
58 can enhance or suppress the mixed layer restratification rate depending on the magnitude
59 and orientation of the wind stress relative to the horizontal density gradient in the ocean
60 mixed layer and the frequency content of the wind [Thomas & Ferrari, 2008; Mahadevan
61 et al., 2010; Long et al., 2012; Whitt et al., 2017a, b] as well as the surface wave field [Li
62 et al., 2012; Haney et al., 2015]. For reviews of recent work on submesoscale impacts on
63 biogeochemistry, see Klein & Lapeyre [2009]; Lévy et al. [2012]; Mahadevan [2016]; Lévy
64 et al. [2018].

65 These conclusions are essentially based on regional and process-study primitive equation
66 ocean models, in which submesoscales are at least partially resolved but small-scale turbu-
67 lence is unresolved and parameterized. It is not clear whether submesoscale interactions
68 with turbulent scales, and therefore the net impacts of submesoscales on biogeochemistry
69 during storms, are realistically represented in such models.

70 Large Eddy Simulations (LES) that explicitly resolve some submesoscales as well as
71 smaller-scale turbulence have started providing additional insights into the interactions

72 between submesoscales and turbulence. For example, such studies have shown energy
73 transfers from submesoscales to small turbulent scales via submesoscale instabilities
74 [e.g., Thomas & Taylor, 2010; Skillingstad & Samelson, 2012; Bachman et al., 2017a;
75 Skillingstad et al., 2017]. If turbulent scale motions are more energetic as a result, this
76 may enhance vertical fluxes. In a recent paper, Whitt & Taylor [2017] used LES to explore
77 the physical response of submesoscales coupled with small-scale turbulence to a storm.
78 Their simulations showed that energetic wind-driven submesoscale motions non-uniformly
79 re-stratified the mixed layer while deepening the reach of boundary-layer turbulence during
80 the storm. This result ensued from spatial heterogeneity in stratification and turbulence.

81 Here, we build on this previous study [Whitt & Taylor, 2017] by describing the in-
82 fluence of submesoscales coupled with small-scale turbulence on biogeochemistry during
83 and shortly after a strong storm using biogeochemical LES (BLES) (for some other re-
84 cent examples of BLES without submesoscales, see Lewis et al. [2017] or Brereton et al.
85 [2018]; for an LES with submesoscales and a plankton model but without nutrients or
86 wind, see Taylor [2016], or for submesoscales and passive tracers under weak atmospheric
87 forcing, see Smith et al. [2016]). Results are obtained by comparing simulations in which
88 submesoscales are present or absent. These results highlight how submesoscale and smaller-
89 scale turbulent nutrient fluxes are modified by the interaction between submesoscale and
90 smaller-scale turbulence, and how this impacts biological production and nutrient and
91 phytoplankton vertical and horizontal distributions. In a second step, we evaluate the ef-
92 ficacy of existing commonly used parameterizations of submesoscales [Fox-Kemper et al.,
93 2008] and small-scale turbulence [Large et al., 1994] in a column model against the BLES
94 benchmarks. We find that the nutrient flux by small-scale turbulence is enhanced by

95 submesoscale interactions during the storm but suppressed shortly after the storm, with
96 the consequence of an increased net biological production, and that the commonly used
97 parameterizations fail at capturing these effects. These results motivate future physi-
98 cal/biogeochemical theory and parameterization development, which can be validated by
99 comparison with these BLES benchmarks and provide a framework for interpreting the
100 simulation results, as well as additional BLES simulations to more fully quantify the sen-
101 sitivity of the simulation results to the parameters that define the numerical experiments.

2. Methods

2.1. Models

102 2.1.1. BLES setup

103 The BLES resolves the majority of the variance associated with three-dimensional fluid
104 motions and is used to evaluate the impact of submesoscale/turbulent-scale interactions
105 on biogeochemistry. The rotating stratified Boussinesq equations are solved on an f -
106 plane without surface wave effects over horizontal wavelengths ranging from about 2 km
107 to about 4 m, which include both submesoscales and smaller-scale turbulence. The nu-
108 merical model, spatial domain and physical initial condition (of density, velocity, and
109 pressure) are as described in Whitt & Taylor [2017] [see also Taylor, 2008]. Briefly, the
110 model solves the Boussinesq equations together with the evolution equations for the four
111 reactive biogeochemical tracers using a low-storage third order Runge-Kutta time march-
112 ing method, a pseudospectral method to represent horizontal derivatives, and second or-
113 der centered differences to represent vertical derivatives. Fluxes of momentum associated
114 with subgrid-scale turbulence with wavelengths smaller than about 4 m are parameter-
115 ized using a modified Smagorinsky model [Kaltenbach et al., 1994]. Subgrid-scale tracer

fluxes also depend on a variable subgrid-scale Prandtl number, which is parameterized in terms of the grid-scale gradient Richardson number as described in Whitt & Taylor [2017]. Subgrid-scale Schmidt numbers associated with biogeochemical tracers are set equal to the subgrid-scale Prandtl number.

Two configurations of the BLES are compared, a frontal zone configuration [e.g., Taylor & Ferrari, 2010; Thomas & Taylor, 2010; Skillingstad et al., 2017; Callies & Ferrari, 2018] (LES,F) with submesoscales, and a configuration without a front (LES,NF) where submesoscales are excluded. The LES,NF configuration captures the conventional effects of small-scale turbulence [e.g., McWilliams & Sullivan, 2000; Skillingstad et al., 2000; Pham & Sarkar, 2017]. Differences between the two configurations quantify how submesoscales modify these conventional effects. In the LES,F configuration, the flow is expressed as a periodic (in x and y) perturbation from a fixed mean horizontal density gradient $\langle M^2 \rangle_{x,y} = \langle \frac{g}{\rho_0} \frac{\partial \rho}{\partial y} \rangle_{x,y} = -\langle \frac{\partial b}{\partial y} \rangle_{x,y} = 5 \times 10^{-8} \text{ s}^{-2}$ and thermal wind shear $\langle M^2 \rangle_{x,y}/f = 5 \times 10^{-4} \text{ s}^{-1}$ that are qualitatively representative of the OSMOSIS site during September of 2012. Here, $\langle \rangle_{x,y}$ denotes a horizontal average, ρ is the density, $\rho_0 = 1026 \text{ kg/m}^3$ is the reference density, g is the acceleration due to gravity, the buoyancy $b = -g\rho/\rho_0$, and $f = 10^{-4} \text{ s}^{-1}$ is the Coriolis frequency. The chosen $\langle M^2 \rangle_{x,y} = 5 \times 10^{-8} \text{ s}^{-2}$ is representative of roughly the 75th percentile of all horizontal buoyancy gradients $|\nabla_h b|$ observed at 2 to 5 km length scales by ocean gliders during autumn at the OSMOSIS site [Thompson et al., 2016]. This means the chosen horizontal density gradient is strong but typical of the region and season; stronger gradients are expected roughly a quarter of the time and weaker gradients are expected roughly three quarters of the time. Hence, although we sometimes refer to the mean horizontal density gradient as a “front,” the

139 gradient is thought to be fairly typical of the open ocean and, in particular, is two orders
140 of magnitude weaker than strong submesoscale fronts that are occasionally observed in
141 boundary currents [*D'Asaro et al.*, 2011].

142 The turbulent physical initial condition at the onset of the storm with finite amplitude
143 submesoscale eddies (Fig. 1) is obtained from a 3-day spin-up simulation that is initialized
144 with low-amplitude red noise in the frontal zone with a vertical density profile based on
145 the observed density profile above the Porcupine Abyssal Plain during September 2012.
146 The spin-up simulation is forced by a constant air-sea buoyancy flux $B_A = 3 \times 10^{-9} \text{ m}^2/\text{s}^3$
147 (equivalent to about $10 \text{ W}/\text{m}^2$ cooling) that is chosen to balance the restratifying effects
148 of submesoscale mixed layer baroclinic instabilities, as discussed in Whitt & Taylor [2017]
149 [see also Mahadevan et al., 2010, 2012]. At the start of the spin-up simulation, the fastest
150 growing mixed layer baroclinic instability has a horizontal length scale of about 985 m
151 and a growth timescale of about 13 hours, as discussed in Whitt & Taylor [2017] [see also
152 Stone, 1966]. The domain size is chosen so that it contains two wavelengths of the fastest
153 growing baroclinic mixed layer instability mode in each horizontal dimension. Hence, the
154 domain is 1970 m by 1970 m by 80 m and covered by a grid with 1024 by 1024 by 160
155 points that achieves a uniform resolution of 1.9 m by 1.9 m by 0.5 m in x (along-front) and
156 y (across-front) and z (vertical), respectively. As a result, the initial perturbations grow
157 and the physical fields have strong submesoscale variability at the onset of the storm, as
158 shown in Fig. 1.

159 The LES,NF configuration is identical to the LES,F configuration except that the do-
160 main is smaller (492.5 m by 492.5 m by 80 m) and $\langle M^2 \rangle_{x,y} = 0$. In particular, this analog
161 configuration uses the same code, the same grid resolution, the same surface boundary

162 conditions, and the same mean density profile $\langle \rho \rangle_{x,y}(z)$ at day 0 as the large domain.
163 But, there is no initial submesoscale variability. Rather, the LES,NF configuration is ini-
164 tialized with low-amplitude white noise in the velocity components and no initial density
165 perturbations.

166 **2.1.2. Storm forcing**

167 The storm forcing is represented by an idealized spatially-uniform but time-dependent
168 surface stress that points 45° to the right of the geostrophic shear; it is inspired by an
169 observed storm during September 24-26, 2012 at the OSMOSIS site [Rumyantseva et al.,
170 2015]. The idealized stress ramps up linearly over two days to a maximum amplitude of
171 0.6 N/m^2 and then ramps down linearly to zero during the third day (as shown in Fig.
172 2). Sensitivity studies with a maximum stress of 0.3 N/m^2 are qualitatively similar and
173 presented in the supplemental material (see Table 1 for a full list of simulations). Unless
174 otherwise noted, the text refers to the cases with the stronger winds. The direction, magni-
175 tude and duration of the wind forcing are chosen to be similar to the observed amplitude,
176 direction and duration of the wind stress during the storm based on NCEP reanalysis
177 winds, satellite sea surface temperature imagery, and results reported in Rumyantseva
178 et al. [2015] and Buckingham et al. [2017].

179 In the presence of a mean horizontal buoyancy gradient at the surface (as in LES,F), the
180 Ekman flow can produce a buoyancy tendency in the mixed layer via horizontal advection.
181 For example, if the wind stress includes a down-front component, meaning a component
182 in the direction of the thermal wind shear $\mathbf{k}/f \times \nabla_h b$, then the Ekman flow tends to
183 reduce the mixed layer buoyancy, energize submesoscale symmetric instabilities aligned
184 parallel to the front, and may cause entrainment. But, if a component of the wind is up

185 front, meaning opposite to the direction of the thermal wind shear, then the Ekman flow
 186 tends to increase the mixed layer buoyancy and may cause subduction [Brannigan, 2016;
 187 Whitt et al., 2017a, b]. In the simulations presented here, the wind includes a down-front
 188 component, and the horizontally-averaged Ekman buoyancy flux $EBF = \tau_x \langle M^2 \rangle_{x,y} / \rho_0 f$,
 189 which is driven by the down-front component of the wind stress τ_x , reaches maximum
 190 magnitudes of $2 \times 10^{-7} \text{ m}^2/\text{s}^3$ (or about $660 \text{ W}/\text{m}^2$ Ekman heat flux) when the magnitude
 191 of the stress is $0.6 \text{ N}/\text{m}^2$ (Fig. 2). It may be noted that the horizontally-averaged EBF is
 192 independent of the interior evolution of the simulations, because the stress is a spatially-
 193 uniform surface boundary condition (and therefore uncorrelated with any interior spatial
 194 variability of density), and the mean horizontal density gradient $\langle M^2 \rangle_{x,y}$ is an externally-
 195 imposed constant.

196 Following the storm, the simulations continue for about 3 days without wind stress. For
 197 simplicity, the air-sea buoyancy flux is held constant at $B_A = 3 \times 10^{-9} \text{ m}^2/\text{s}^3$ (or about
 198 $10 \text{ W}/\text{m}^2$ heat loss) during and after the simulated storm.

199 It may be noted that the simulations presented here do not have realistic air-sea buoy-
 200 ancy loss, which is usually enhanced during storms. That is because, although we have run
 201 simulations with more realistic buoyancy loss (not shown), the stronger buoyancy loss has
 202 little impact on the results. A wind stress of $0.6 \text{ N}/\text{m}^2$ is a far stronger source of energy for
 203 turbulent entrainment in a 40 m deep mixed layer than the roughly $200 \text{ W}/\text{m}^2$ net heat
 204 loss observed by Rumyantseva et al. [2015] (as can be inferred from scaling arguments
 205 [Kraus and Turner, 1967; Pollard et al., 1972] and observations [Thomas et al., 2016]).
 206 Further, even a realistic surface buoyancy flux would be about a factor of 3 smaller than
 207 the EBF. Using the observationally-supported scalings reported in Thomas et al. [2016],

208 the depth-integrated dissipation of turbulent kinetic energy should have contributions of
 209 about $6 \times 10^{-5} \text{ m}^3/\text{s}^3$ from wind and $8 \times 10^{-6} \text{ m}^3/\text{s}^3$ from EBF during peak wind. That
 210 is to say, the direct injection of energy by the wind during the storm is the dominant
 211 source of energy for the turbulence. Based on these scalings, the turbulence dynamics
 212 are not expected to differ much between the various scenarios presented here, nor is the
 213 biogeochemical response.

214 2.1.3. Biogeochemical model

215 The biogeochemical model is a modified version of the NPZD (nutrient, phytoplankton,
 216 zooplankton, detritus) model implemented by Whitt et al. [2017b] and Whitt et al. [2017a],
 217 which is based on classic models like that of Fasham et al. [1990] and a previous NPZD
 218 implementation by Powell et al. [2006]. The biogeochemical model equations are:

$$\begin{aligned}
 219 \quad \frac{DN}{Dt} &= -UP + \gamma_n GZ + \delta D + \nabla \cdot ((\kappa_{SGS} + \kappa_0) \nabla N) - \beta(N - N_{80}), \\
 220 \quad \frac{DP}{Dt} &= UP - \sigma_d P - GZ + \nabla \cdot ((\kappa_{SGS} + \kappa_0) \nabla P), \\
 221 \quad \frac{DZ}{Dt} &= (1 - \gamma_n)GZ - \zeta Z - \hat{\zeta} Z^2 + \nabla \cdot ((\kappa_{SGS} + \kappa_0) \nabla Z), \\
 222 \quad \frac{DD}{Dt} &= \sigma_d P + \zeta Z + \hat{\zeta} Z^2 - \delta D + w_d \frac{\partial D}{\partial z} + \nabla \cdot ((\kappa_{SGS} + \kappa_0) \nabla D), \\
 223 \quad G &= R(1 - e^{-\Lambda P}), \\
 224 \quad U &= \frac{V_m N}{k_N + N} \frac{\alpha I}{\sqrt{V_m^2 + \alpha^2 I^2}}, \\
 225 \quad I &= I_0 \exp(k_w z),
 \end{aligned} \tag{1}$$

226 where $D/Dt = \partial/\partial t + \mathbf{u} \cdot \nabla$ is the material derivative, the subgrid-scale diffusivity is given
 227 by κ_{SGS} (which varies in time and space), the molecular diffusivity is set to a constant
 228 $\kappa_0 = 10^{-6} \text{ m}^2/\text{s}$, and the constant parameters are listed in Table 2.

229 The biogeochemical initial conditions (Fig. 3), which are imposed as horizontally-
 230 uniform fields at $t = 0$ d, are from an equilibrium (i.e., steady state) solution to the
 231 biogeochemical model equations with the parameters in Table 2 and a constant verti-
 232 cal diffusivity $\kappa_{SGS} + \kappa_0 = 2 \times 10^{-5}$ m²/s. The initial Z and D profiles (not shown)
 233 have a similar vertical structure to P , with maximum concentrations of 0.22 and 0.35
 234 mmol/m³ respectively. This equilibrium solution is calculated numerically by running
 235 one-dimensional simulations of the biogeochemical model equations for 10 years with a
 236 one-dimensional NPZD model described earlier [Whitt, 2017; Whitt et al., 2017b, a]. In
 237 addition, numerical experimentation suggests that the resulting equilibrium profile is ef-
 238 fectively independent of the initial conditions of N , P , Z , and D for the parameter set in
 239 Table 2.

240 Due to the construction of the biogeochemical initial condition, the horizontally-
 241 averaged nutrient $\langle N \rangle_{x,y}$ and phytoplankton $\langle P \rangle_{x,y}$ profiles have strong vertical gradients
 242 throughout the top 60 m at day 0 (Fig. 3). In the upper 30 m, these strong vertical bio-
 243 geochemical gradients contrast with weak vertical density gradients, which are relatively
 244 permissive of turbulent mixing compared to the pycnocline below. Hence, shortly after
 245 the simulations begin, storm-driven turbulence drives rapid vertical mixing and homoge-
 246 nization of the mean biogeochemical concentrations over the top 30 m.

247 To reflect the idealized nature of the simulations, we refer to the simulated nutrient as
 248 “nutrient” throughout the paper. In addition, it is not our intention to validate the model
 249 or exactly reproduce observations here. However, the biogeochemical model parameters
 250 in Table 2 are tuned within reasonable ranges (see the appendix) so that the chosen
 251 initial equilibrium, which reflects the parameters, has a nutrient profile that is similar to

252 the observed dissolved silica profile in autumn 2012 above the Porcupine Abyssal Plain on
253 September 21-22, 2012 a few days before the storm (Fig. 3 (b)). The subsurface maximum
254 in the initial phytoplankton profile is also qualitatively consistent with observations of
255 vertical profiles of chlorophyll fluorescence (Fig. 3 (a)).

256 **2.1.4. Column set up in ROMS**

257 In order to assess the effectiveness of some existing ocean model parameterizations at
258 representing the horizontally-averaged behavior of LES, results from the BLES models are
259 compared to results from a one-dimensional column model implemented in the regional
260 ocean modeling system (ROMS+NPZD) [e.g., Powell et al., 2006; Whitt et al., 2017b].
261 Two ROMS configurations are used that can be viewed as parameterized analogues of
262 the two LES configurations, one with parameterized submesoscales (ROMS,F) and one
263 with no submesoscales (ROMS,NF). In the ROMS,NF configuration, the physical effects
264 of turbulence are parametrized using the conventional KPP mixing scheme [Large et al.,
265 1994], as implemented in the public version of ROMS (myroms.org). There are no effects
266 of submesoscales and no horizontal density gradient, i.e. $\langle M^2 \rangle_{x,y} = 0$. In the ROMS,F
267 configuration, the physical effects of both turbulence and submesoscales are parametrized
268 rather than explicitly resolved, and the same mean horizontal density gradient as in the
269 LES,F experiments is used, i.e. $\langle M^2 \rangle_{x,y} = 5 \times 10^{-8} \text{ s}^{-2}$.

270 In all column model experiments, ROMS solves for the time evolution of horizontal
271 momentum, temperature (which is equivalent to density via a linear equation of state),
272 and biogeochemical tracers on the same uniform vertical grid with 160 levels and 0.5
273 m resolution that is used in the LES. Vertical fluxes are computed using a conservative

274 parabolic spline method. In addition, the column model is forced by the same surface
 275 momentum and buoyancy fluxes and solves the same biogeochemical model equations.

276 The parameterizations for small-scale turbulence and for submesoscales are implemented
 277 in the column version of ROMS as follows. First, the KPP surface layer scheme is modified
 278 so that the boundary layer depth, and thereby the diffusivity and viscosity profiles, are
 279 defined based on a bulk Richardson number that depends on the constant geostrophic
 280 shear $\langle M^2 \rangle_{x,y}/f$ as well as the evolving ageostrophic shear. Second, terms are added to
 281 the right side of the temperature equation (which is also the density equation since we
 282 are using linear equation of state) to represent horizontal advection of the background
 283 lateral temperature gradient, including an extra term of the form $v\langle M^2 \rangle_{x,y}/g\alpha$ where α is
 284 a constant thermal expansion coefficient. This extra advection term includes the Ekman
 285 buoyancy flux EBF defined earlier. In addition, a second term is added to the right hand
 286 side of the temperature equation of the form $v^*\langle M^2 \rangle_{x,y}/g\alpha$, where v^* is defined below.
 287 This second new term in the temperature equation represents the restratifying effect of
 288 submesoscale mixed layer baroclinic instabilities via the parameterization of Fox-Kemper
 289 et al. [2008] (FFH), in which the temperature tendency associated with this restratifying
 290 effect is expressed as $-\nabla \cdot (\mathbf{u}^*T)$, where

$$291 \quad v^* = -C \frac{MLD_3^2 \langle M^2 \rangle_{x,y}}{|f|} \frac{\partial \mu(z)}{\partial z} \quad (2)$$

292 and the other components of the bolus velocity vector \mathbf{u}^* are equal to zero in this case.

293 The time-dependent mixed layer depth MLD_3 in (2) is defined in section 2.2.2 below.

294 The time-dependent vertical structure function $\mu(z)$, which depends on MLD_3 as well as
 295 z , is given by (39) of FFH. The coefficient $C = 0.06$, which FFH find to be the best fit in a
 296 series of simulations without diurnal convection or wind, but this parameter C is subject

297 to some uncertainty [e.g., Bachman & Taylor, 2016; Taylor, 2016; Callies & Ferrari, 2018].
298 Finally, the convergence of the viscous vertical flux of geostrophic momentum is added
299 to the tendency. This flux of geostrophic momentum is calculated like the viscous flux
300 of ageostrophic momentum, by multiplying the same vertically and temporally variable
301 viscosity coefficients to the constant background geostrophic shear profile.

2.2. Model diagnostics

2.2.1. Submesoscales and smaller-scale turbulence: scale separation

302
303 Throughout this paper, the term submesoscale is used descriptively to refer to fea-
304 tures of the velocity, density, and tracer fields that reside primarily in the ocean mixed
305 layer with characteristic horizontal length scales of 0.1-10 km, and, unless stated other-
306 wise, does not imply any particular physical process. In the context considered in this
307 paper—the mid-latitude North Atlantic during autumn—this definition implies that sub-
308 mesoscales have horizontal scales that are larger than the mixed layer depth, which is
309 generally 10-100 m. Therefore, submesoscales have small aspect ratios and are distinct
310 from conventional mixed layer turbulence, which is typically composed of features with
311 horizontal wavelengths of the same order as or smaller than the mixed layer depth (i.e.,
312 $\lesssim 0.1$ km horizontal scales) and order-one aspect ratios. Since some simulations reveal
313 a change in the slope of kinetic energy and tracer spectra at a wavelength slightly larger
314 than the mixed layer depth [Hamlington et al., 2014; Smith et al., 2016; Whitt & Taylor,
315 2017], we define a separation at the horizontal wavelength of 150 m to separate smaller
316 turbulence from larger submesoscales in the computation of the variances and vertical
317 fluxes [the rationale behind this particular choice is discussed in Whitt & Taylor, 2017].
318 For example, the total vertical nutrient flux $\int \hat{w}(k)\hat{N}^*(k) dk$ is separated into its subme-

319 soscale and turbulent parts, using a separating wavenumber 1/150 cyc/m. The results
 320 presented here do not change qualitatively if this separation wavenumber is increased or
 321 decreased by a factor of two. Hence, we say the LES,F scenarios in the larger 2 km domain
 322 contain submesoscales, and we say the LES,NF scenarios in the smaller 0.5 km domain
 323 do not contain submesoscales (see Table 1 for a list of simulations).

324 2.2.2. Mixed-layer depths

325 Two different density thresholds are used to define two mixed-layer depths [see Brain-
 326 erd & Gregg, 1995]: The first MLD_{05} is defined by a relatively small density threshold
 327 $\Delta\langle\rho\rangle_{x,y} = 0.005 \text{ kg/m}^3$ to approximately capture the subtle effect of intermittent strat-
 328 ification within the turbulent boundary layer. Conversely, the second MLD_3 is defined
 329 by a more conventional density threshold $\Delta\langle\rho\rangle_{x,y} = 0.03 \text{ kg/m}^3$ to approximately capture
 330 the depth that intermittent boundary-layer turbulence reaches. Unless otherwise noted,
 331 “mixed layer depth” or “mixed layer base” refers to the more conventional MLD_3 .

332 2.2.3. Net community production

333 Net community production NCP is equal to production by autotrophs minus respiration
 334 by the entire community over a given time period. In this model, $NCP = \int UP -$
 335 $\gamma_n GZ - \delta D dt$, that is the time integral of the reaction terms on the right hand side
 336 of the N evolution equation. NCP has not been diagnosed online during the model
 337 integration and was computed offline from changes in N . More precisely, the reduction in
 338 full-column-integrated $\langle N \rangle_{x,y}$ during the storm is equal to the storm-driven net community
 339 production $\langle NCP \rangle_{x,y}$. Indeed, if the N evolution equation is integrated over the entire
 340 model domain, then the advective and diffusive terms go to zero. In addition, since the
 341 nutrient concentration at the bottom grid cell of the domain (80 m) is not modified by the

342 storm, the nutrient restoring tendency at the bottom grid cell does not induce a change in
 343 full-column-integrated N during the storm. Hence, any simulated changes in full-column-
 344 integrated $\langle N \rangle_{x,y}$ during the storm are caused by the reactive tendency terms in the N
 345 equation, i.e. $\langle NCP \rangle_{x,y}$. Unless otherwise noted, column-integrated henceforth implies
 346 integration over the full model domain depth, i.e. $H = 80$ m.

347 **2.2.4. Primary productivity eddy reactions**

Both nutrient variability as well as nutrient covariability with phytoplankton have a
 rectified impact on primary productivity $\langle UP \rangle_{x,y}$. This effect is due to correlations in the
 biogeochemical concentrations below the domain scale, e.g. $\langle N'P' \rangle_{x,y}$ where the primes
 denote deviations from the horizontal average [see Lévy & Martin, 2013; Martin et al.,
 2015; Woodson & Litvin, 2015]. It may be noted that the leading order (quadratic) terms
 of the eddy reactions in the mean primary productivity $\langle UP \rangle_{x,y}$ are proportional to (by
 Taylor's approximation):

$$\frac{\langle N'P' \rangle_{x,y}}{k_N + \langle N \rangle_{x,y}} - \frac{\langle N'P' \rangle_{x,y} \langle N \rangle_{x,y} + \langle N'N' \rangle_{x,y} \langle P \rangle_{x,y}}{(k_N + \langle N \rangle_{x,y})^2} + \frac{\langle N'N' \rangle_{x,y} \langle N \rangle_{x,y} \langle P \rangle_{x,y}}{(k_N + \langle N \rangle_{x,y})^3} \quad (3)$$

348 where the primes denote deviations from the lateral average and it is assumed that
 349 $|N'|/(k_N + \langle N \rangle_{x,y}) \ll 1$. In the LES scenarios presented in this paper, where
 350 $\langle P \rangle_{x,y} \lesssim k_N + \langle N \rangle_{x,y}$, the magnitude of the variances $var(N)$ and $var(P)$, where
 351 $var(N) = \langle N'N' \rangle_{x,y}$, can be used to bound the magnitude of the ratio of primary produc-
 352 tivity eddy reactions to the part of $\langle UP \rangle_{x,y}$ associated with the mean, which is proportional
 353 to $\langle N \rangle_{x,y} \langle P \rangle_{x,y} / (k_N + \langle N \rangle_{x,y})$. In particular, the ratios are bounded by products of the
 354 coefficients of variation (e.g., $c_v(N)c_v(P)$ where $c_v(N) = \sqrt{var(N)_{x,y}} / \langle N \rangle_{x,y}$), by the
 355 Cauchy-Schwartz inequality.

356 **2.2.5. Potential energy change due to vertical mixing**

357 The potential energy ($PE = \int \rho g(z + H) dz$ where z ranges from -80 m to 0 m and
 358 $H = 80$ m is the domain depth [e.g., Crawford & Large, 1996; Burchard and Hofmeister,
 359 2008]) can be increased in three ways in response to the storm. First, vertical mixing acts
 360 to raise the center of mass and thereby increases the potential energy. Second, vertically-
 361 sheared horizontal mean currents can act on the mean horizontal density gradient to
 362 reduce the vertical stratification and hence also increase the potential energy. For wind-
 363 driven (Ekman) currents, the corresponding buoyancy flux is referred to as an Ekman
 364 buoyancy flux (EBF). Third, PE can increase due to a stabilizing air-sea buoyancy flux
 365 (e.g., surface heating). See Whitt & Taylor [2017] for more details on the energetics.

In order to quantify the approximate increase in PE due to mixing, the change in PE due to lateral advection and the air-sea buoyancy flux, B_A , must be subtracted from the change in PE, ΔPE , that is

$$\Delta PE^* = \Delta PE - \left(\Delta \int_{-D}^0 \rho g dz \right) \frac{1}{MLD_3} \int_{-MLD_3}^0 (z + H) dz. \quad (4)$$

366 Dividing the residual change in potential energy by the wind work $\Delta PE^*/\Delta KE$, where
 367 $\Delta KE = \int \boldsymbol{\tau} \cdot \mathbf{u} dt$ and \mathbf{u} is the surface ocean velocity, approximately quantifies the fraction
 368 of the wind work that goes toward increasing the potential energy via turbulent mixing,
 369 i.e., a mixing ratio $r_m = \Delta PE^*/\Delta KE$.

370 2.2.6. Vertical diffusivity

The diffusivities in ROMS are a direct output of the model/KPP scheme. In the LES, the vertical diffusivities are estimated by adding the subgrid-scale contribution to the resolved vertical nutrient flux and dividing the result by the mean vertical nutrient gradient,

$$\kappa_N = \underbrace{\frac{-\langle wN \rangle_{x,y}}{\frac{\partial \langle N \rangle_{x,y}}{\partial z}}}_{\text{resolved}} + \underbrace{\frac{\langle (\kappa_{SGS} + \kappa_0) \frac{\partial N}{\partial z} \rangle_{x,y}}{\frac{\partial \langle N \rangle_{x,y}}{\partial z}}}_{\text{SGS model}}. \quad (5)$$

3. Results

3.1. Mean properties of the LES solutions

371 The first question addressed in the results is: can the fairly typical horizontal den-
372 sity gradient and submesoscale heterogeneity in the mixed layer impact the mean phys-
373 ical/biogeochemical tracer response in this strongly forced scenario, in which the storm
374 winds are the dominant source of energy to the boundary layer? And, if so, by how
375 much? To contextualize the biogeochemical response to the storm, we first describe the
376 influence of the front and submesoscales on the evolution of the mixed layer depth and
377 stratification during the scenario. A description of how stratification evolves in the LES,F
378 and LES,NF experiments with the weaker wind can be found in Whitt & Taylor [2017],
379 and the stronger wind scenario presented here is qualitatively similar (the supplemental
380 material contains figures showing comparisons between the simulations with stronger and
381 weaker wind stresses, which are enumerated in Table 1).

3.1.1. Density

383 At the onset of the storm, density is already fairly well mixed in the upper 30 m (Fig. 3).
384 Yet, the storm is sufficiently strong to mix denser water up from the pycnocline, to deepen
385 MLD_3 by approximately 10 m and to increase the surface density (Figs. 4 (c) and (f)).
386 The deepening of MLD_3 is approximately the same in the scenarios with and without
387 submesoscales, but the MLD_3 deepens roughly half as much when the wind stress is half
388 as strong (suppl. Fig. 5). That is, the deepening of MLD_3 is insensitive to the horizontal
389 density gradient or the presence of submesoscale heterogeneity, up to the degree each
390 are present in these simulations. However, this is not the case for MLD_{05} , which shoals
391 during the storm with submesoscales and deepens without submesoscales (Fig. 4; see also

392 suppl. Fig. 5). The MLD_{05} is shallower in the simulation with submesoscales because the
 393 vertical density gradient above MLD_3 is sensitive to the presence of the submesoscales
 394 during the storm. Without submesoscales, the vertical density gradient above MLD_3
 395 is concentrated very close to MLD_3 , whereas with submesoscales the vertical density
 396 gradient above MLD_3 is distributed more evenly with depth. Hence, the curvature of
 397 the vertical density profile is weaker near MLD_3 with submesoscales, which is consistent
 398 with enhanced vertical mixing of density across MLD_3 during the storm in that case. We
 399 can also note that the surface density is greater after the storm with submesoscales, but
 400 this denser surface water could be attributable to the horizontal Ekman buoyancy flux
 401 and/or enhanced vertical transport from the pycnocline. Therefore, the denser surface
 402 water with submesoscales is not necessarily associated with increased vertical mixing.

403 In order to show that the submesoscale heterogeneity triggers enhanced vertical ex-
 404 change of density from the pycnocline to the mixed layer during the storm, we quantify
 405 the increase in potential energy due to vertical mixing alone (ΔPE^* , defined in section
 406 2.2.5). We find that ΔPE^* is about twice as large with submesoscales as without (Fig.
 407 5), and we obtain a quantitatively similar result for both wind stress magnitudes (suppl.
 408 Fig. 6). The mixing ratio ($r_m = \Delta PE^*/\Delta KE$, defined in section 2.2.5) ranges from
 409 about 1.5% without submesoscales to 3% with submesoscales (Fig. 5 (b)), and r_m is only
 410 reduced by about 1/4 when the stress is half as strong, although both ΔPE^* and the
 411 kinetic energy input by the wind ΔKE are reduced by about 2/3 with the weaker stress
 412 (suppl. Fig. 6). These results show that the submesoscales enhance the net (subme-
 413 soscale+turbulent) vertical transport of density across the mixed layer base relative to

414 the simulations without submesoscales, without significantly increasing the kinetic energy
415 input by the wind.

416 Before discussing the biogeochemical response, it is worth noting that a parameteri-
417 zation of r_m is a crucial explicit or implicit choice (e.g., via a bulk Richardson number
418 threshold, as discussed in *Pollard et al.* [1972]) in many models of ocean boundary layer
419 mixing and stratification [*Kraus and Turner*, 1967; *Pollard et al.*, 1972; Large et al., 1994;
420 Crawford & Large, 1996; *Burchard and Hofmeister*, 2008], and these choices can strongly
421 impact the modeled response of the mixed layer to wind forcing. However, this is the first
422 time that LES results have been used to explicitly show that submesoscale heterogeneity
423 can systematically enhance r_m , and no mixing scheme that is used operationally in ocean
424 models includes a parameter to explicitly enhance the mixing efficiency to account for
425 unresolved submesoscale effects like those simulated here. Based on this new result, it is
426 to be expected that submesoscales also enhance the mean vertical nutrient flux $\langle wN \rangle_{x,y}$
427 and net community production $\langle NCP \rangle_{x,y}$ in this nutrient-limited autumn storm scenario.
428 In what follows, we quantify and explore these biogeochemical impacts in more detail.

429 3.1.2. Nutrient and phytoplankton

430 We start by examining how the horizontally-averaged phytoplankton and nutrient pro-
431 files evolve during and after the storm in the LES,F and LES,NF simulations. By some
432 measures, the mean biogeochemical response to the storm is similar with and without
433 submesoscales (Fig. 4). Storm-driven turbulence fluxes nutrients and phytoplankton up
434 towards higher light levels near the ocean surface. The storm-driven upward biogeochem-
435 ical fluxes trigger a reduction in column-integrated nutrient $\langle N \rangle_{x,y}$ and a nearly equal
436 and opposite increase in column-integrated phytoplankton biomass $\langle P \rangle_{x,y}$ (Fig. 6). This

437 mirroring of the changes in column-integrated $\langle N \rangle_{x,y}$ and $\langle P \rangle_{x,y}$ indicates that primary
438 productivity is the dominant mechanism of column-integrated biogeochemical transforma-
439 tion. These results are qualitatively similar under the weaker stress, but the changes in
440 column-integrated $\langle N \rangle_{x,y}$ and $\langle P \rangle_{x,y}$ are about 1/3 to 1/2 as large, which is qualitatively
441 consistent with less deepening of the MLD_3 and less nutrient entrainment under weaker
442 wind stress (suppl. Fig. 5). In addition, this evolution is qualitatively consistent with
443 expectations based on available observations of the nutrient and/or chlorophyll responses
444 to autumn storm passage [e.g., Babin et al., 2004; Rumyantseva et al., 2015].

445 However, the mean biogeochemical response to the storm also reveals some systematic
446 differences with and without submesoscales. First, submesoscales enhance the column-
447 integrated $\langle NCP \rangle_{x,y}$ by about a factor of two, which is qualitatively consistent with the
448 increased mixing ratio r_m . In addition, the column-integrated $\langle NCP \rangle_{x,y}$ is also enhanced
449 by submesoscales when the wind is half as strong, albeit by somewhat less than a factor
450 of two (suppl. Fig. 5). We can also note a reduced curvature of the mean vertical
451 nutrient and phytoplankton profiles near MLD_3 in the scenario with submesoscales, which
452 is consistent with the reduced curvature of the density profiles near MLD_3 and enhanced
453 vertical mixing there with submesoscales. Finally we also find faster surface accumulation
454 of phytoplankton and stronger vertical gradients in both phytoplankton and nutrient
455 that develop within the mixed layer in the scenarios with submesoscales compared to the
456 scenarios without submesoscales. In the following section, we show how submesoscales
457 modulate the net (submesoscale+turbulent) vertical fluxes of biogeochemical constituents
458 to explain these results.

3.2. Horizontal variability in the LES solutions

459 Based on the results of the previous section, it is expected that average vertical nutri-
 460 ent fluxes $\langle wN \rangle_{x,y}$ will be sensitive to the presence of submesoscales. However, since the
 461 biogeochemical dynamics, such as primary productivity and grazing, are non-linear, the
 462 biogeochemical covariability at scales from meters to submesoscales, such as $cov(N, P)$,
 463 may rectify on larger-scale biogeochemical dynamics as well. In addition, the fluxes and
 464 covariances are best understood in the context of the variances, such as $var(N)$, on which
 465 they depend. Hence, this section addresses the following questions for the first time with
 466 BLES: does submesoscale heterogeneity impact the horizontal biogeochemical variances,
 467 such as $var(N)$, or key covariances, such as $cov(N, P)$, which impacts primary productiv-
 468 ity, or $cov(w, N) = \langle wN \rangle_{x,y}$, in this storm scenario? To what degree do the submesoscales
 469 (> 150 m) and turbulent scales (< 150 m) contribute to the total variances and covari-
 470 ances during the different phases of the storm? Is there a rectified effect of submesoscales
 471 on physical and biogeochemical variances and covariances at turbulent scales?

472 We find that the physical/biogeochemical variances and covariances are greatly modified
 473 by the presence of submesoscales during and after the storm. To gain further perspective
 474 on this aspect of the solutions, we first qualitatively compare the lateral variability of
 475 the nutrient N and nutrient flux wN in snapshots from the different solutions during the
 476 storm (Fig. 7) before proceeding to statistical measures of the variability. Just above
 477 MLD_3 (i.e., $MLD_3 + 4$ m), the variability in nutrient concentration differs qualitatively
 478 between LES,F and LES,NF. With submesoscales, the nutrient distribution reflects the
 479 submesoscale vertical displacements of the nutricline (Figs. 7 (a)), which dominate the
 480 total variability. However, the dominant characteristic scales of the nutrient flux wN

481 (which are qualitatively similar to w , not shown) are much smaller. Yet, these dominant
 482 small-scale structures in wN are modulated somewhat by the larger submesoscales (Fig.
 483 7 (b)). In contrast, the LES,NF scenario exhibits no obvious qualitative evidence of
 484 organization at submesoscales in either N or wN (Fig. 7), and the characteristic horizontal
 485 scales of N and wN variability are comparable in magnitude to MLD_3 . In this sense,
 486 the simulated variability of N and wN in the LES,NF scenario is qualitatively consistent
 487 with classic ideas about boundary layer turbulence, in which the largest aspect-ratio-one
 488 turbulent eddies with a length scale similar to MLD_3 dominate both the variances and
 489 the fluxes. In contrast, the LES,F scenario is qualitatively inconsistent with classic ideas
 490 about boundary layer turbulence in the same sense, because low-aspect-ratio structures
 491 that are a factor of 10-50 times larger than MLD_3 qualitatively dominate the variance.
 492 (For a spectral perspective on this statement, compare the schematic power spectrum of
 493 boundary layer turbulence in Fig. 1 of *Wynngaard* [2004], in which the power drops at
 494 horizontal wavelengths greater than the boundary layer height, to simulated power spectra
 495 in Fig. 5 of *Whitt & Taylor* [2017], in which the power rises at horizontal wavelengths
 496 greater than the mixed layer depth).

497 **3.2.1. Biogeochemical variances in the LES solutions**

498 For a more statistical perspective on the dependence of the biogeochemical variability on
 499 the submesoscales, we explicitly quantify and compare the vertical profiles of the nutrient
 500 variance $var(N)$, and we show how submesoscales (>150 m) and smaller turbulent scales
 501 (<150 m) contribute to the total variance in the LES,F and LES,NF scenarios (Fig. 8).
 502 The phytoplankton variance profiles $var(P)$ are qualitatively similar to $var(N)$, and the

503 results are qualitatively insensitive to reducing the wind stress by half (see suppl. Figs.
504 7-8), so we only discuss the stronger wind scenario in this context.

505 During the storm (days 2-3), $var(N)$ is maximal just below MLD_3 and declines mono-
506 tonically upwards so that it is 3-4 orders of magnitude smaller at the surface. With
507 submesoscales, there is much greater total variance in the mixed layer due almost entirely
508 to submesoscale (> 150 m) variance, similar to weakly-forced LES simulations of passive
509 tracers released in the pycnocline by Smith et al. [2016]. However, in a more novel re-
510 sult, the presence of submesoscales also increases the variance of N at smaller (< 150 m)
511 turbulent scales during the storm by up to about an order of magnitude, particularly in
512 the upper part of the mixed layer (Fig. 8 (C)). The enhanced smaller-scale N variance
513 with submesoscales is likely due primarily to enhanced physical variance production from
514 stronger mean and submesoscale vertical gradients (Fig. 4), since the smaller-scale tur-
515 bulent kinetic energy and the dissipation of turbulent kinetic energy in the mixed layer
516 are only modestly modified by submesoscales during the storm (see suppl. Fig. 12 here
517 and Fig. 5 of Whitt & Taylor [2017]).

518 Although kinetic energy at submesoscales and turbulent scales declines substantially
519 after the storm [Whitt & Taylor, 2017] (see also suppl. Fig. 12), $var(N)$ generally increases
520 after the storm and remains about 1-2 orders of magnitude greater with submesoscales
521 than without. With submesoscales, $var(N)$ achieves its maximum in the middle of the
522 mixed layer, i.e. from about 10 to 30 m depth (Fig. 8 (F)), due to a substantial increase
523 in submesoscale variance at these depths after the storm. This increase in submesoscale
524 variance after the storm is likely due primarily to the near shutdown of turbulent mixing
525 and the corresponding reduction of the rate of dissipation of submesoscale N variance

526 via turbulent mixing. On the other hand, production of submesoscale variance is not
 527 substantially enhanced after the storm, since submesoscale kinetic energy declines after
 528 the storm [Whitt & Taylor, 2017], and mean vertical tracer gradients are only slightly
 529 enhanced with submesoscales (Fig. 4).

530 3.2.2. Biogeochemical covariances in the LES solutions

531 The rectified impacts of nutrient and phytoplankton variability on primary productiv-
 532 ity, which is defined by $\langle UP \rangle_{x,y}$ in (1), require spatial correlation, $corr(N, P)$ in addition
 533 to variance (see section 2.2.4). In the mixed layer, $corr(N, P)$ is time and depth depen-
 534 dent; it is positive during the storm and negative after the storm, but mostly order-one
 535 (see suppl. Fig. 9). Therefore, eddy reactions are non-zero and larger in magnitude
 536 with submesoscales due to the larger horizontal variances, and the eddy reactions can
 537 take both signs. However, it may be noted that the smallness of both coefficients of
 538 variation $c_v(N)$ and $c_v(P)$ directly imply the smallness of primary productivity eddy re-
 539 actions relative to the part of $\langle UP \rangle_{x,y}$ associated with the means, which is proportional to
 540 $\langle N \rangle_{x,y} \langle P \rangle_{x,y} / (k_N + \langle N \rangle_{x,y})$ (see section 2.2.4). In addition, the storm-driven perturbation
 541 to zooplankton production is small compared to the perturbation to primary production
 542 in this scenario. So, we do not pursue questions about eddy reactions any further here,
 543 although the dynamic nature of the variances of N and P and their correlation reflect non-
 544 trivial physical/biogeochemical dynamics that have never been previously documented in
 545 BLES, and future work is needed to fully investigate these dynamics in this and other
 546 scenarios.

547 3.2.3. Nutrient fluxes in the LES solutions

548 Here, we quantify and discuss the vertical nutrient flux, and its submesoscale and tur-
549 bulent parts in the LES,F and LES,NF scenarios (Fig. 9; $var(N)$ is in Fig. 8; $var(w)$ is
550 in suppl. Fig. 12; and multi-scale correlation profiles $corr(w, N)$ are in suppl. Fig. 13).
551 Consistent with the results above, we find that both the total fluxes and their respec-
552 tive submesoscale and turbulent parts differ by an order-one fraction with and without
553 submesoscales, both during and after the storm.

554 First, we find that the presence of submesoscales substantially enhances the total vertical
555 nutrient flux during the storm (between days 2-3, left column in Fig. 9). In addition, the
556 effect of the submesoscales on the maximum total flux is qualitatively similar but greater
557 in the scenario with stronger stress (a 115% increase) than the scenario with the weaker
558 stress (a 34% increase, see suppl. Fig. 10). However, the total flux profiles exhibit a
559 qualitatively similar vertical structure in scenarios with and without submesoscales. In
560 both scenarios, the maximum flux occurs between about $0.8MLD_3$ and $0.9MLD_3$, and
561 the flux drops almost linearly to zero between the depth where it achieves its maximum
562 and the surface above or about $1.1MLD_3$ below. Yet, the strong fluxes penetrate deeper
563 below the mixed layer MLD_3 in the LES,F scenarios compared to the LES,NF scenarios,
564 which is consistent with deeper penetration of high rates of dissipation of turbulent kinetic
565 energy [Whitt & Taylor, 2017] and vertical velocity variance (suppl. Fig. 12) into the
566 pycnocline/nutricline in the presence of submesoscales.

567 The partitioning of the flux between submesoscales (> 150 m) and smaller turbulent
568 scales (< 150 m) during the storm differs significantly depending on the presence or
569 absence of the submesoscales, and the differences are qualitatively similar with weaker
570 wind (suppl. Fig. 10). Without submesoscales in LES,NF, the total flux is dominated

571 by smaller turbulent scales. On the other hand, with submesoscales in LES,F, subme-
572 soscale and smaller-scale turbulent fluxes are both positive and submesoscales contribute
573 an order-one fraction of the total flux. The strong submesoscale flux is qualitatively con-
574 sistent with strong submesoscale N variance (Fig. 8), and it shows that the submesoscale
575 heaving of the nutricline observed in Fig. 7 is also associated with some irreversible up-
576 ward nutrient transport at submesoscales. With a weaker wind, the fraction of the total
577 flux from submesoscales is larger, but the magnitude of the submesoscale flux is smaller
578 (see suppl. Fig. 10). Incidentally, the decrease in submesoscale flux with the decrease
579 in wind stress magnitude is qualitatively consistent with previous process studies with
580 ocean models that parameterize small turbulent scales [e.g., Mahadevan & Tandon, 2006;
581 Capet et al., 2008a; Thomas et al., 2008; Whitt et al., 2017b], although the causes of there
582 relationships are not necessarily the same (see Whitt & Taylor [2017] for some relevant
583 discussion). However, in a more novel result, the smaller-scale turbulent contribution to
584 the nutrient flux is significantly greater during the storm with submesoscales than with-
585 out, even though vertical velocity variance is only modestly altered throughout most of
586 the mixed layer (suppl. Fig. 12). This enhanced flux (i.e., covariance) is qualitatively
587 consistent with enhanced turbulent N variance with submesoscales, but the correlation
588 between nutrient and vertical velocity is actually smaller, both in aggregate and at tur-
589 bulent scales, in the simulations with submesoscales than without submesoscales (suppl.
590 Fig. 13). Finally, perhaps the most intriguing feature of the nutrient flux profiles during
591 the storm is the deeper penetration of the strong vertical fluxes below the mixed layer in
592 the LES,F scenarios with submesoscales compared to the LES,NF scenarios without. This
593 enhanced flux near MLD_3 occurs in conjunction with a substantial increase in smaller-

594 scale turbulent vertical velocity variance (suppl. Fig. 12) and is due mostly although not
595 entirely to stronger smaller-scale fluxes (Fig. 9).

596 After the storm, the nutrient flux profiles change dramatically compared to their coun-
597 terparts during the storm. In particular, the magnitude of the total flux is reduced by
598 about an order of magnitude in both simulations (note the different x-axes in the left
599 and right columns in Figure 9). In addition, the maximum total flux is about a factor of
600 two weaker with submesoscales than without. The partitioning of the total flux between
601 submesoscales and turbulent scales is also very different between the two scenarios. In
602 particular, the smaller-scale turbulent flux is much weaker in the scenario with the sub-
603 mesoscales (LES,F) than in the scenario without submesoscales (LES,NF). Consequently,
604 after the storm the total flux is achieved mostly by the submesoscale flux in LES,F and
605 by the smaller-scale turbulent flux in LES,NF.

606 The shut-down of the turbulent flux after the storm in the LES,F scenario and cor-
607 responding increase in vertical biogeochemical gradients is qualitatively consistent with
608 earlier results showing how submesoscales influence the timing of the spring bloom by en-
609 hancing the density stratification, suppressing turbulence, and thereby modifying vertical
610 fluxes in deeper spring mixed layers [Taylor, 2016]. More generally, these results suggest
611 that fronts/submesoscales may contribute to the rapid formation of biogeochemical gradi-
612 ents within mixed layers after strong wind events during any time of year by suppressing
613 turbulent fluxes even when the mixed layer depths have not yet shoaled significantly [e.g.,
614 as observed by Carranza et al., 2018].

3.3. Comparisons between the LES and column model mean solutions

615 The comparison between the LES results of the previous section (LES,F and LES,NF)
 616 and two analog column model simulations (ROMS,F and ROMS, NF) is presented with
 617 the aim of evaluating the ability of existing ocean model parameterizations (encapsulated
 618 in the column model) to plausibly represent the physical-biogeochemical dynamics sim-
 619 ulated by LES. We focus mostly on the scenarios with the front/submesoscales, which
 620 are the most interesting, but nevertheless show the results for the scenarios without a
 621 front/submesoscales as a reference. For the two models (LES and ROMS), we compare
 622 the total flux and total diffusivities (Figs. 10-11). Comparison of the mean density,
 623 nutrient, phytoplankton profiles are shown in suppl. Figs. 2-4. Comparisons of the
 624 column-integrated quantities, such as column integrated $\langle N \rangle_{x,y}$ and $\langle P \rangle_{x,y}$ as well as the
 625 change in potential energy and the mixing ratio are in suppl. Figs. 5-6. In addition,
 626 fluxes and diffusivities in scenarios with the weaker wind, which are qualitatively similar
 627 to their counterparts in the scenarios with stronger wind, are shown in suppl. Figs. 10-11.

628 3.3.1. Scenarios without a front or submesoscales

629 Broadly speaking, the column model ROMS,NF is rather effective at reproducing the
 630 laterally-averaged evolution in the LES,NF simulation without a front/submesoscales by
 631 all the metrics considered. Vertical tracer fluxes in ROMS,NF are a fairly good repre-
 632 sentation of vertical tracer fluxes in LES,NF. The ROMS,NF scenario also produces very
 633 similar mixed-layer depths and vertical diffusivity profiles during and after the storm
 634 compared to the LES,NF scenario (Fig. 11). As a consequence, the mean profiles and
 635 column integrals of phytoplankton and nutrient are also similar, under both weak and
 636 strong wind (Figs. 12 (c)-(d); see supplementary Figs. 2-3 for the profile comparisons).

637 This result confirms previous findings [e.g., Large et al., 1994; Large & Crawford, 1995;
638 Skillingstad et al., 2000; McWilliams & Sullivan, 2000; Pham & Sarkar, 2017] that the
639 KPP parameterization is capable of capturing the change in vertical mixing in response
640 to a storm in the absence of a front/submesoscales.

641 3.3.2. Scenarios with a front and submesoscales

642 The frontal zone version of the column model ROMS,F does not perform as well
643 at capturing the physical/biogeochemical dynamics of the LES,F scenario with the
644 front/submesoscales. In fact, the ROMS,F scenarios are in many ways more similar to
645 the ROMS,NF and LES,NF scenarios than their analog LES,F scenarios, which shows
646 that the submesoscale heterogeneity simulated by LES,F is the cause of the differences
647 relative to LES,NF, not the front. As in previous sections, only the scenarios forced
648 by the stronger winds are presented here; the scenarios forced by the weaker winds are
649 qualitatively similar and appear in suppl. Figs. 2-6 and 10-11.

650 The most distinctive result emerging from the LES,F/LES,NF comparison above is that
651 the presence of submesoscales in LES,F roughly doubles the vertical nutrient fluxes and
652 mixing ratio during the storm, and the submesoscales indirectly cause turbulence and
653 turbulent fluxes to penetrate below MLD_3 . In contrast, the total vertical nutrient flux
654 in the ROMS,F scenario is actually reduced compared to the ROMS,NF and LES,NF
655 scenarios during the storm. As a result, the maximum vertical nutrient flux in ROMS,F
656 is about a quarter of that in LES,F (Fig. 10 (a)). Further, ROMS,F fails to capture the
657 deeper penetration of vertical fluxes below MLD_3 that is simulated in LES,F.

658 Comparing the vertical diffusivity profiles κ_N during the storm (Fig. 10 (c)) shows
659 that ROMS,F fails to capture the strong vertical nutrient fluxes in the mixed layer partly

660 because the mean nutrient gradients are too weak, not just because the diffusivities are too
661 weak. For example, between about $0.5MLD_3$ and $0.9MLD_3$, the diffusivity is actually
662 lower in LES,F than ROMS,F, while the fluxes are about 3-4 times larger in LES,F
663 than ROMS,F (mean nutrient profiles are compared in suppl. Fig. 3). On the other
664 hand, just below the mixed layer (between about MLD_3 and $1.1MLD_3$), both the flux
665 and diffusivity are much larger in the LES,F than ROMS,F. These results suggest that
666 submesoscales systematically modify the empirical non-dimensional shape functions as
667 well as the turbulent boundary layer depth, which characterize the diffusivity profiles in
668 the mixing scheme [Large et al., 1994], during the storm.

669 After the storm, there is some indication that the physics encapsulated in the ROMS,F
670 scenario may capture some of the essential physics in the LES,F scenario. In particular, the
671 maximum total fluxes in the mixed layer are smaller in magnitude and occur at a shallower
672 depth than in the corresponding LES,NF and ROMS,NF scenarios (c.f., Figs. 10 (b) and
673 11 (b)). In addition, both the mixed layer depth MLD_{05} and the surface mixing layer
674 depth shoal in the ROMS,F scenario. However, there are still some discrepancies between
675 LES,F and ROMS,F. For example, the maximum nutrient flux is only about a quarter to
676 a third of that in LES,F. In the upper half of the mixed layer, where vertical diffusivities
677 in ROMS,F are mostly larger than in LES,F, the reduced vertical fluxes are caused by
678 too-weak mean gradients. However, between $0.5MLD_3$ and MLD_3 , the reduced vertical
679 fluxes are mostly caused by too-small vertical diffusivities. In addition, the reduced fluxes
680 are presumably partially a result of missing submesoscale vertical fluxes, which dominate
681 in LES,F after the storm, but are not included in ROMS,F.

682 Despite differences, these significantly reduced diffusivities after the storm—in both
683 ROMS,F and LES,F—manifest in enhanced mean vertical gradients in phytoplankton
684 above MLD_3 by day 5-6 in the scenarios with a front compared to their counterparts
685 without a front, that is LES,NF and ROMS,NF (suppl. Figs. 2-3). Hence, the relatively
686 rapid formation of biogeochemical vertical gradients within the mixed layer in both LES,F
687 and ROMS,F scenarios is partly attributable to the substantial reduction in vertical mixing
688 in the scenarios in a frontal zone. However, as a result of the failure of the ROMS,F column
689 model to capture the enhanced vertical nutrient fluxes during the storm, there is only a
690 third to a half as much net community production and accumulation of column-integrated
691 phytoplankton during and shortly after the storm (Fig. 12 (b)).

692 **3.3.3. Summary of the comparisons between the column model and LES**

693 Qualitatively and quantitatively, the conventional boundary layer column model
694 ROMS,NF accurately captures the physical-biogeochemical evolution of the conventional
695 boundary layer LES,NF without a front/submesoscales. This is encouraging, with the
696 caveat that there is some indication that the performance of KPP may not be equally
697 good in other parts of physical parameter space [Pham & Sarkar, 2017]. Regardless, the
698 column model with a frontal zone ROMS,F fails to reproduce the physical-biogeochemical
699 evolution in the LES with a frontal zone. In particular, the column model fails to cap-
700 ture the fact that submesoscale physical variability in the mixed layer enhances vertical
701 nutrient fluxes, phytoplankton accumulation, and net community production during the
702 life cycle of the storm. However, the column model does capture the qualitatively reduced
703 mixing within the mixed layer MLD_3 and the related rapid emergence of vertical phy-

704 top plankton gradients within the mixed layer after the storm with a front/submesoscales
705 compared to scenarios without a front/submesoscales.

706 The failure of the column model to capture the enhanced vertical mixing and nutrient
707 fluxes during the storm rules out several a priori plausible explanations for this differ-
708 ence between LES,F and LES,NF, including the Ekman buoyancy flux, vertical mixing
709 of geostrophic momentum, and/or the restratifying buoyancy flux associated with mixed
710 layer baroclinic instabilities as described by FFH via (2). However, these three factors
711 could contribute indirectly, in combination with other properties of submesoscale vari-
712 ability in LES,F, to induce enhanced mixing during the storm. On the other hand, the
713 partial success of the column model in capturing the reduced vertical mixing and the more
714 rapid formation of vertical gradients in phytoplankton after the storm suggests that the
715 column model captures some of the essential physical-biogeochemical dynamics associated
716 with the rapid formation of these vertical phytoplankton gradients in LES,F. In partic-
717 ular, the combination of the Ekman buoyancy flux, mixing of geostrophic momentum,
718 and restratification via (2) results in substantially reduced mixing within the mixed layer
719 after the storm, and the more rapid formation of biogeochemical gradients within the
720 mixed layer. Future work is required to generalize these results and to develop and test
721 other hypotheses about the key physical-biogeochemical mechanisms responsible for the
722 enhanced vertical mixing and net community production during the storm.

4. Conclusion and Discussion

4.1. Summary

723 This study uses process-oriented numerical simulations to address the questions
724 of whether and to what degree submesoscale heterogeneity influences the physical-

725 biogeochemical response to a storm. Previous process studies using numerical models
726 with parameterized turbulence have demonstrated that submesoscales have a significant
727 impact on mixed layer tracer budgets and hence biogeochemical dynamics, including and
728 perhaps especially under wind forcing [e.g., Franks & Walstad, 1997; Mahadevan et al.,
729 2008; Lévy et al., 2009; Brannigan, 2016; Whitt et al., 2017b]. However, the results
730 presented here highlight and address an important source of uncertainty affecting these
731 previous studies: the turbulence parameterizations have neither been designed for nor
732 validated in an ocean surface mixed layer with significant submesoscale heterogeneity.

733 The results presented here are an important contribution because they are derived from
734 coupled physical-biogeochemical ocean simulations that explicitly resolve a significant
735 range of both submesoscales and turbulent scales and thereby greatly reduce the uncer-
736 tainty emanating from turbulence parameterizations. In particular, the LES presented
737 here are all in the high-Peclet number regime, that is $Pe \approx 50$ meaning that the resolved
738 turbulent advective vertical tracer flux is about 50 times stronger than the subgrid-scale
739 diffusive vertical tracer flux in the mixed layer during the storm (suppl. Fig. 11). Hence,
740 the qualitative results are expected to be insensitive to further refinement of the grid
741 resolution or changes to the subgrid-scale model. This expectation is supported by a grid
742 refinement study (not shown), in which the first 1.7 days of the LES,NF scenario were
743 re-run in a smaller domain (320 m wide), and the entrainment was effectively identical
744 on grids with horizontal resolutions ranging from 2.2 m to 1.1 m, which span the 1.9 m
745 resolution used for all the simulations reported above.

746 The results presented here answer the scientific question set forth in the affirma-
747 tive: submesoscale heterogeneity in a frontal zone significantly modifies the physical-

748 biogeochemical response to the storm compared to otherwise identical scenarios without
749 the submesoscales. Most importantly, submesoscales strongly enhance the wind-driven
750 flux of tracers from the pycnocline to the mixed layer. For example, submesoscales en-
751 hance the maximum vertical nutrient flux in the mixed layer by a factor of two and the
752 turbulent diffusivity at the mixed layer base by a factor of ten. Interestingly, the decom-
753 position of the vertical nutrient fluxes into larger submesoscales and smaller turbulent
754 scales shows that submesoscales contribute to enhanced vertical fluxes during the storm
755 in two ways. First, the vertical fluxes associated with submesoscale features directly trans-
756 port nutrients from the pycnocline to the mixed layer, as observed in previous studies.
757 However, these results also show that submesoscale physical variability in a frontal zone
758 acts as a catalyst by enhancing the small turbulent-scale nutrient fluxes near and just be-
759 low the mixed layer base. On the other hand, submesoscales suppress net vertical tracer
760 fluxes after the storm by inducing a near shut-down of turbulent scale fluxes, although
761 submesoscale fluxes actually increase after the storm. Hence, submesoscales facilitate a
762 more rapid emergence of vertical phytoplankton gradients within the mixed layer after
763 the storm. Consequently, mean phytoplankton accumulation rates are roughly twice as
764 fast after the storm in the scenarios with submesoscales compared to their counterparts
765 without.

766 Since the column model simulations do not capture this enhanced mean phytoplank-
767 ton accumulation rate in the frontal zone configuration, existing physical-biogeochemical
768 ocean models with horizontal grid spacings of 2 km or greater and the KPP and FFH
769 parameterizations will not effectively capture the enhanced storm-driven phytoplankton
770 growth simulated in LES. In addition, these results demonstrate that the combination of

771 processes added to the column model with a frontal zone, which include the destratifying
772 Ekman buoyancy flux and vertical mixing of geostrophic momentum that would also be
773 included in a submesoscale-permitting regional ocean model, cannot explain the enhanced
774 fluxes and biogeochemical impacts on their own. Rather, other processes or combinations
775 of processes must be invoked to explain the results. However, the significance of these
776 results is far from fully understood at this point, because these results are derived from
777 only a few case studies. In addition, since the underlying processes that explain these
778 results are not fully understood, they may impact other properties of the upper ocean
779 beyond biogeochemistry, including heat, salt, momentum, and kinetic energy budgets.

4.2. Comparison with other process simulations

780 Although this is the first study to present LES that resolve substantial parts of the
781 submesoscale range together with a fully-coupled biogeochemical model (with coupled
782 nutrient/ecosystem dynamics), the results are consistent with and build on a few previous
783 studies that have presented LES that resolve both submesoscales and turbulence and
784 simulate passive tracers and/or phytoplankton.

785 For example, the result that submesoscales enhance the lateral biogeochemical variance
786 by an order of magnitude or two after the storm is qualitatively consistent with the LES
787 simulations of passive tracers in Smith et al. [2016], where it is shown that submesoscales
788 enhance the lateral variance of tracers initialized near the base of the mixed layer by up
789 to a factor of 20 under weak winds and weak cooling. However, the results presented
790 here show that submesoscales significantly enhance nutrient and phytoplankton variance
791 during the strong wind forcing for the first time, although the submesoscale tracer variance
792 in the mixed layer is generally lower during the storm than afterwards. In addition, these

793 results show that submesoscales indirectly enhance smaller-scale turbulent tracer variance
 794 during the storm.

795 In addition, the result that the submesoscales are associated with enhanced wind-driven
 796 vertical mixing near the mixed layer base is qualitatively consistent with several earlier
 797 LES in short/wide frontal zones, which have shown that strong turbulent vertical mixing
 798 of tracers extends to deeper depths than predicted by KPP [Bachman et al., 2017a].
 799 However, these earlier results are attributed to symmetric instability turbulence mixing
 800 over the depth of the low potential vorticity layer H_q (defined to be the deepest depth
 801 where $\int_{H_q}^0 f \langle \frac{\partial b}{\partial z} \rangle_{x,y} + \langle \frac{\partial u}{\partial z} \rangle_{x,y} \langle \frac{\partial b}{\partial y} \rangle_{x,y} dz = 0$; see also Whitt et al. [2017b]). In the LES
 802 simulations presented here, enhanced vertical mixing extends well below H_q and even
 803 penetrates into the high stratification of the seasonal pycnocline below the mixed layer
 804 [Whitt & Taylor, 2017]. In addition, these results show explicitly that the enhanced
 805 vertical fluxes with submesoscales are due in large part to enhanced turbulent scale fluxes,
 806 in addition to enhanced submesoscale vertical fluxes.

807 Finally, the result that submesoscales reduce net vertical fluxes after the storm is also
 808 qualitatively consistent with the conclusions of Smith et al. [2016], in which it is shown
 809 that submesoscales inhibit wind/wave driven vertical mixing of passive tracers released
 810 near the mixed layer base under weak atmospheric forcing. In addition, the results are
 811 consistent with the emerging hypothesis that submesoscales suppress net vertical fluxes
 812 under weak forcing (e.g., after the storm), but submesoscales do not suppress [as in Taylor,
 813 2016] or enhance (as in this case) net vertical fluxes as well as smaller-scale turbulent fluxes
 814 under sufficiently strong forcing.

4.3. Comparison with observations

815 It is difficult to observe all the variables and spatio-temporal scales necessary to evaluate
816 the hypotheses emerging from these simulations, but there are many observations of the
817 bio-optical and physical response of the upper-ocean to a summer or autumn storm from
818 satellites [e.g., Lin et al., 2003; Babin et al., 2004; *Fauchereau et al.*, 2011; *Lin*, 2012;
819 Carranza & Gille, 2015] and a rapidly increasing but small number of subsurface bio-
820 optical observations during and after storms from profiling subsurface floats [e.g., *Ye et al.*,
821 2013; *Chacko*, 2017; *Girishkumar et al.*, 2019]. Like all our simulations, many of these
822 prior observations, including the observations of Rumyantseva et al. [2015] and Painter
823 et al. [2016], which motivated the present study, show that autumn storms increase the
824 surface density, deepen the mixed layer, erode subsurface chlorophyll maxima and mix
825 subsurface chlorophyll up toward the surface (thereby increasing the fraction of column-
826 integrated chlorophyll in the mixed layer), and trigger an increase in column-integrated
827 chlorophyll. However, the magnitude and qualitative nature of these responses are highly
828 variable between different storms [see e.g., Painter et al., 2016] due to the wide variety of
829 different physical and biogeochemical circumstances in which storms occur.

830 Although prior observations reveal mesoscale spatial variability in the surface temper-
831 ature and chlorophyll response to autumn storms [e.g., Babin et al., 2004; *Lin*, 2012;
832 Painter et al., 2016; *Girishkumar et al.*, 2019], most observations do not provide explicit
833 information about the mesoscale lateral density variability in conjunction with explicit
834 information about the corresponding lateral variability of submesoscales, turbulence and
835 biogeochemistry during and after a storm. Without such data, it is difficult to observa-
836 tionally evaluate the hypotheses emerging from the LES, specifically that even relatively

837 modest open-ocean mesoscale fronts (in comparison to areas without fronts) are associ-
838 ated with: 1) enhanced submesoscale and turbulent-scale nutrient fluxes during a storm,
839 2) more intermittent turbulence and suppressed net turbulent nutrient fluxes in the mixed
840 layer after the storm, and 3) enhanced storm-driven new production.

841 Nevertheless, we can consider these hypotheses in light of some rare, nearly-simultaneous
842 observations of turbulence, nutrients and submesoscales. For example, the OSMOSIS
843 program, which motivated these process simulations, obtained observations of turbulent
844 nutrient fluxes and submesoscales before and after an autumn storm. In addition, the
845 OSMOSIS program provides broader context for these focused process studies, since re-
846 sults are available showing the full seasonal cycle of bio-optical parameters like chlorophyll
847 fluorescence, backscatter, among others [*Erickson and Thompson, 2018; Bol et al., 2018*].
848 The observations during the storm at the OSMOSIS site reveal highly-intermittent tur-
849 bulent nutrient fluxes near the mixed layer base at a single location for one day after
850 the storm. Although Rumyantseva et al. [2015] find that the intermittent mixing is plau-
851 sibly attributable to intermittent alignment between the ocean currents and the wind,
852 our simulations with a front and submesoscales (LES,F) also reveal large spatio-temporal
853 intermittency in turbulent nutrient fluxes near the mixed layer base, both during and
854 after the storm even though the effective wind speed after the storm is zero. Figure 7 (b)
855 shows how nutrient fluxes are modulated by submesoscales during the simulated storm;
856 qualitatively similar results are found for turbulent nutrient fluxes after the storm, but
857 not shown; see also Whitt & Taylor [2017]. In addition, the hypothesis that submesoscale
858 variability could have contributed to the observed intermittency of the turbulent nutrient
859 flux in the observations of Rumyantseva et al. [2015] is bolstered by satellite observations

860 of substantial submesoscale variability in the sea surface temperature before the storm
861 [Buckingham et al., 2017]. However, subsurface observations of scales smaller than 2 km
862 are not available.

863 Elsewhere, a high-resolution physical/biogeochemical survey in the vicinity of the
864 Kuroshio front by Nagai & Clayton [2017] reveals submesoscale lateral variability in nutri-
865 ent concentrations, which is qualitatively similar to the simulated submesoscale nutrient
866 variability near the pycnocline in Figure 7 (a). In addition, Nagai & Clayton [2017] find
867 that these nutrient anomalies have a rectified effect on the net turbulent nutrient flux,
868 consistent with our LES results. However, their observations were not during/after a
869 storm.

870 A future interdisciplinary observational study is needed to evaluate the hypothesis that
871 mesoscale oceanic lateral density variability induces lateral variations in the submesoscale
872 and turbulent-scale physical and biogeochemical response to a storm using multi-scale
873 sampling of nutrients (in various forms), turbulence, and primary productivity during
874 the life cycle of a storm at resolutions down to ~ 1 km over a region $\lesssim 100$ km in scale.
875 Such a dataset is challenging to obtain due to the operational challenge associated with
876 collecting water samples during a storm, but it may be possible with an array of robotic
877 towed and/or autonomous samplers.

4.4. Remaining uncertainties associated with the model configurations

878 Finally, it is worth reiterating that the results of these simulations should be viewed
879 with caution due to a variety of outstanding uncertainties that have not been addressed
880 here. Perhaps most significantly, the physics may be influenced by the small domain
881 size and the absence of larger submesoscale and mesoscale structures. In addition, the

882 biogeochemical model is at best a qualitative representation of the real biogeochemical
883 dynamics, and it is expected that the biogeochemical model may influence the details of
884 the results. Further, we have only one initial condition and two fairly-idealized forcing
885 scenarios, which omit surface wave effects.

886 If general results are to be obtained, then future work must explore sensitivities to the
887 details of the forcing, including wind stress direction and frequency content, as well as air-
888 sea buoyancy flux and surface wave effects. With regard to waves, real storms are expected
889 to generate strong wind waves, which substantially modify the Lagrangian mean shear in
890 the upper ocean. Hence, as several previous studies have shown, waves can substantially
891 modify the upper-ocean turbulence [e.g., McWilliams & Sullivan, 2000; Li et al., 2012],
892 submesoscale/turbulence interactions [Hamlington et al., 2014; Haney et al., 2015; Smith
893 et al., 2016], and thereby vertical tracer transport. It is difficult to say how waves might
894 impact the results in this scenario, because no previous study has considered the impacts
895 of waves on submesoscale/turbulence interactions under strong storm winds, as in this
896 scenario. However, we expect that the addition of wind waves, roughly aligned with the
897 winds, would be an additional source of energy for smaller-scale turbulence and enhance
898 entrainment. But, it is less clear how these waves would impact the relatively strong flow
899 of kinetic energy to submesoscales during the storm in these simulations. Hence, a high
900 priority for future work should be to explore submesoscale/turbulence interactions with
901 strong waves and strong winds, building on studies with weak wind [e.g., Hamlington
902 et al., 2014; Smith et al., 2016].

903 In addition, future work will have to explore the sensitivity of the results to the strength
904 of the initial horizontal and vertical buoyancy gradients, the initial submesoscale pertur-

905 bations, as well as the domain size and the presence/absence of larger scale structures.
906 Future work should also explore the sensitivity to the biogeochemical parameters and
907 model formulation.

908 In summary, the results presented above demonstrate that submesoscale heterogeneity
909 can modify the mixing-induced increase in potential energy and thereby the biogeochem-
910 ical response to a storm by order-one fractions. These results are novel and because they
911 are from explicit coupled physical-biogeochemical large eddy simulations, which are the
912 first to couple an NPZD model to LES. In addition, comparisons between the LES and a
913 vertical column ocean model show for the first time that submesoscale/turbulence inter-
914 actions, which are not captured by current theory and ocean model parameterizations but
915 simulated by LES, are responsible for the enhanced storm-driven biogeochemical response.
916 Thus, submesoscale/turbulence interactions that are not represented in ocean/Earth sys-
917 tem models may modify phytoplankton seasonal phenology in storm-driven mid-latitude
918 oceans. Since the biogeochemical impacts of submesoscale/turbulent scale physics in these
919 idealized experiments is rather strong, future work is needed to assess the broader sig-
920 nificance, understand the sensitivity of the results to various parameters and values not
921 considered here, and parameterize these cross-scale physical interactions as well as the
922 combined physical-biogeochemical interactions.

923 Thus, further experiments to address outstanding uncertainties with expensive
924 submesoscale-and-turbulence resolving BLES are justified. Nevertheless, a thorough ex-
925 ploration of these sensitivities using LES may not be feasible. A promising alternative
926 direction is to use high-resolution LES like those presented here selectively to facilitate
927 the development and validation of ocean models with horizontal resolutions between 10-

100 m, which can be used to broadly explore the parameter space at a reasonable cost if they perform well in comparisons with LES. However, even this latter approach may require unconventional and expensive non-hydrostatic physical/biogeochemical ocean models. Nevertheless, non-hydrostatic ocean models at 10-100 m resolution are 2-4 orders of magnitude cheaper to run than non-hydrostatic LES at ~ 1 m resolution in the same domain. In addition, some studies have used hydrostatic ocean models with horizontal resolutions down to about 100 m, although the validity of hydrostatic ocean model solutions of the ocean boundary layer at the scales below 2 km is not well understood. Thus, an important outstanding question is if and to what degree a hydrostatic ocean model with ~ 100 m resolution, and therefore only a 10 by 10 or 20 by 20 grid over the 2 km box and a turbulence parameterization, can represent the dynamics simulated using BLES in this paper.

Appendix: Rationale behind biogeochemical model parameters

The final set of biogeochemical model parameters in Table 2 was determined from many trial simulations of biogeochemical equilibrium profiles with different biogeochemical parameters but the same constant vertical diffusivity using a one-dimensional NPZD model Whitt [2017]. The approach to identifying the final parameter set in Table 2 was ad hoc; parameters are constrained to be reasonably consistent with published data, prior model studies focusing on the eastern North Atlantic, and via inspection of the resulting equilibrium modeled profiles in comparison with the observations reported in Rumyantseva et al. [2015] and Painter et al. [2016] and other available observations. The final parameter set in Table 2 was chosen primarily by identifying (by eye) equilibrium vertical profiles of nutrient (N) that matched the observed dissolved silica (DIS) profile reasonably well (Fig.

3). The chosen parameters yield an initial profile that achieves this goal. Although more
rigorous optimization is certainly possible, such an effort is beyond the scope of this paper
and in our opinion is unlikely to yield qualitatively different results in the experiments
presented here. The results are derived from comparisons between different physical mod-
els with the same biogeochemical model and don't depend on the model being able to
exactly reproduce observations as long as it is qualitatively reasonable. A rationale for
some of the chosen parameters based on available data and previously published literature
is reported below.

The surface photosynthetically available radiation (PAR) is set to $I_0 = 50 \text{ W/m}^2$, which
is equal to the climatological PAR from the pixel nearest the study site (48.7° N , 16.4° W)
averaged between September and October as obtained from the VIIRS instrument [Ocean
Ecology Laboratory, Ocean Biology Processing Group NASA Goddard Space Flight Cen-
ter., 2018]. The e-folding depth scale for the attenuation of light with depth is set to a
constant $1/k_w = 18.2 \text{ m}$, which is intermediate between the observed e-folding depth of
about 10 m [Rumyantseva et al., 2015] and the e-folding depth in clear water in the region
of about 30 m [Fasham & Evans, 1995]. For the parameters in Table 2, the compensation
depth where phytoplankton photosynthesis is equal to phytoplankton cellular respiration,
that is where $U = \sigma_d$, and the 1% light level are both between 80 and 90 m . The initial
equilibrium profiles are highly sensitive to the chosen e-folding depth $1/k_w$. If the other
parameters are held fixed, neither the e-folding depth chosen by Fasham & Evans [1995]
nor the value observed by Rumyantseva et al. [2015] produces an equilibrium profile that
is a good match to observations before the storm. For simplicity, phytoplankton have
no explicit influence on the PAR profile, unlike most other NPZD models. Rumyantseva

973 et al. [2015] observe relatively small temporal changes in the e-folding depth of the PAR
974 profile during the storm event, so the dynamic nature of the PAR profile is thought to be
975 a minor effect.

976 Thee phytoplankton parameters, α , V_m , and k_N , and σ_d were taken from the models
977 of Fasham & Evans [1995], and are fairly typical of NPZD-class models. The mortality
978 parameter σ_d only indirectly affects the solution (via its influence on the initial condition)
979 in these short 6-day experiments, which are about a quarter of the phytoplankton specific
980 mortality e-folding timescale. The zooplankton parameters, R , Λ , γ_n , ζ and $\hat{\zeta}$ are less well
981 known than the phytoplankton parameters and are tuned somewhat to produce reasonable
982 initial conditions, but these are still fairly typical for NPZD-class models.

983 The detrital remineralization depth scale (sinking velocity w_d divided by specific rem-
984 ineralization rate δ) in the models of Fasham & Evans [1995] ranges from 100 to 400 m.
985 Here, a much smaller value of 2.5 m is used in order to get an equilibrium in the 80 m
986 depth domain with such simplified forcing (i.e., with a constant vertical diffusivity and no
987 surface mixing layer). This is mostly achieved by both reducing the sinking speed from
988 3-6 m/d to $w_d = 0.25$ m/d and increasing the detrital remineralization rate from 0.01-
989 0.06 d^{-1} to 0.1 d^{-1} . These choices help in achieving realistic nutrient and phytoplankton
990 profiles in a one-dimensional equilibrium without seasonal nutrient replenishment by deep
991 mixing. The chosen sinking velocity is somewhat unconventional and therefore should be
992 viewed largely as a tuning knob that helps to create an initial equilibrium profile that is
993 similar to observations in this study. This is a reasonable approach in this case, because
994 remineralization timescales are generally longer than the timescales of interest here (<6
995 days) and neither export nor remineralization are explicitly considered in this paper.

1000 In order to crudely represent all external physical sources of nutrient that are missing
1001 in this model, such as upwelling or vertical mixing of water with higher nutrient concen-
1002 trations from deeper depths, nutrient is restored to a value of $N_{80} = 2.5 \text{ mmol/m}^3$ over a
1003 timescale $1/\beta = 4$ days at the bottom grid cell only, that is at 80 m depth. The objective
1004 of this restoring is primarily to achieve an equilibrium vertical nutrient profile like the
1005 observed DIS profile observed during September 2012 (Fig. 3 (b)). The choice to tune
1006 the model parameters to fit the observed DIS profile as opposed to other macronutrients
1007 probably does not change the qualitative results very much. Observed dissolved inorganic
1008 nitrogen (DIN) and phosphorous (DIP) profiles look much the same as DIS, hence initial-
1009 izing with those profiles would also result in an enhanced vertical nutrient flux during the
1010 storm. However, the observed stoichiometric ratios are not perfectly linear (see Fig. 13 of
1011 Painter et al. [2016]). For example, the DIN:DIS ratio is about 5:1 at small concentrations
1012 in the upper 60 m but follows a 2:1 line at higher concentrations between 60 m and 120
1013 m. In addition, the DIN:DIP ratio follows an 18:1 line. The ratios of particulate organic
1014 nitrogen (PON) to biogenic silica (BSi) are about 5:1 to 10:1 in the upper 50 m [see Fig.
1015 14 of Painter et al., 2016].

1016 The quantitative results presented in this paper are sensitive to the chosen biogeochem-
1017 ical parameters, which are reflected in the one-dimensional equilibrium solution used as
1018 the initial condition for the forced experiments. However, it is beyond the scope of this
1019 paper to show explicitly how either the biogeochemical initial conditions or the results
1020 of the wind-forced experiments are sensitive to changes in the biogeochemical model pa-
1021 rameters. The qualitative and quantitative sensitivity of one-dimensional NPZD-class
1022 models to changes in the biogeochemical model parameters is discussed elsewhere [e.g.,

1019 McGillicuddy et al., 1995; Doney et al., 1996; Evans & Garçon, 1997; Franks, 2002; Beck-
1020 mann & Hense, 2007]; recent and relevant discussions about the sensitivity of the phyto-
1021 plankton and nutrient responses to wind-driven nutrient entrainment events in a similar
1022 NPZD-class model can be found in Whitt et al. [2017b] and Whitt et al. [2017a] (n.b.,
1023 supplemental section 4.1 of Whitt et al. [2017b]). The upshot is that the quantitative
1024 biogeochemical results should not be viewed as an exact reproduction of specific ocean
1025 conditions, despite some efforts documented above to choose parameters that are relevant
1026 to the observations reported by Rumyantseva et al. [2015] and Painter et al. [2016] in
1027 particular and the eastern North Atlantic more generally. However, the modeled biogeo-
1028 chemical responses to the different transient forcing scenarios in different physical models,
1029 all with the same biogeochemical model equations, parameters, and initial conditions,
1030 provide perspective on the physics of the different scenarios as well as the qualitative
1031 sensitivity of the biogeochemical response to physical differences alone. Hence, the dis-
1032 cussion of the results focuses on the qualitative features of the biogeochemical solutions
1033 and comparisons between scenarios over the precise quantitative features of the results,
1034 which can nevertheless be obtained from the figures and the public data [Whitt, 2018].

Acknowledgements

1035 DBW thanks Genevieve Brett for helpful discussions on an earlier version of the
1036 manuscript. DBW was supported by the National Science Foundation, via an NSF
1037 Postdoctoral Fellowship (OCE-1421125) as well as NSF grants OPP-1501993, and OCE-
1038 1658541, as well as NOAA via NA18OAR4310408. ML acknowledges support from ANR
1039 SOBUMS (ANR-16-CE01-0014) and CNES. JRT was supported by the Natural Envi-
1040 ronment Research Council, via award NE/J010472/1. We would also like to acknowl-

1041 edge high-performance computing support from Yellowstone ([ark:/85065/d7wd3xhc](https://doi.org/10.5065/d7wd3xhc)) and
1042 Cheyenne ([doi:10.5065/D6RX99HX](https://doi.org/10.5065/D6RX99HX)) provided by NCAR’s Computational and Informa-
1043 tion Systems Laboratory, sponsored by the National Science Foundation. The Na-
1044 tional Center for Atmospheric Research is sponsored by the National Science Foundation.
1045 The relevant source code for DIABLO (github.com/danielwhitt/DIABLO) and ROMS
1046 (github.com/danielwhitt/ROMS) is publicly available, and the initial condition data nec-
1047 essary to reproduce the simulation is archived online [Whitt, 2018]. The plotting scripts
1048 and associated data are archived with the initial conditions.

References

- 1049 Babin, S., Carton, J., Dickey, T., & Wiggert, J. (2004). Satellite evidence of hurricane-
1050 induced phytoplankton blooms in an oceanic desert. *Journal of Geophysical Research:*
1051 *Oceans*, *109*.
- 1052 Bachman, S. D., Fox-Kemper, B., Taylor, J. R., & Thomas, L. (2017a). Parameterization
1053 of frontal symmetric instabilities. i: Theory for resolved fronts. *Ocean Modelling*, *109*,
1054 72–95.
- 1055 Bachman, S. D., & Taylor, J. R. (2016). Numerical simulations of the equilibrium between
1056 eddy-induced restratification and vertical mixing. *Journal of Physical Oceanography*,
1057 *46*, 919–935.
- 1058 Bachman, S. D., Taylor, J. R., Adams, K., & Hosegood, P. (2017b). Mesoscale and
1059 submesoscale effects on mixed layer depth in the southern ocean. *Journal of Physical*
1060 *Oceanography*, *47*, 2173–2188.

- 1061 Beckmann, A., & Hense, I. (2007). Beneath the surface: Characteristics of oceanic ecosys-
1062 tems under weak mixing conditions—a theoretical investigation. *Progress in Oceanogra-*
1063 *phy*, *75*, 771–796.
- 1064 Boccaletti, G., Ferrari, R., & Fox-Kemper, B. (2007). Mixed layer instabilities and re-
1065 stratification. *J. Phys. Oceanogr.*, *37*, 2228–2250.
- 1066 Bol, R., S. A. Henson, A. Rumyantseva, and N. Briggs (2018), High-frequency variability
1067 of small-particle carbon export flux in the northeast atlantic, *Global biogeochemical*
1068 *cycles*, *32*(12), 1803–1814.
- 1069 Brainerd, K. E., & Gregg, M. C. (1995). Surface mixed and mixing layer depths. *Deep*
1070 *Sea Research Part I: Oceanographic Research Papers*, *42*, 1521–1543.
- 1071 Brannigan, L. (2016). Intense submesoscale upwelling in anticyclonic eddies. *Geophys.*
1072 *Res. Lett.*, . doi:10.1002/2016GL067926. 2016GL067926.
- 1073 Brereton, A., Siddons, J., & Lewis, D. (2018). Large-eddy simulation of subsurface phyto-
1074 plankton dynamics: an optimum condition for chlorophyll patchiness induced by Lang-
1075 muir circulations. *Marine Ecology Progress Series*, *593*, 15–27.
- 1076 Buckingham, C. E., Khaleel, Z., Lazar, A., Martin, A. P., Allen, J. T., Naveira Garabato,
1077 A. C., Thompson, A. F., & Vic, C. (2017). Testing munk’s hypothesis for subme-
1078 soscale eddy generation using observations in the north atlantic. *Journal of Geophysical*
1079 *Research: Oceans*, .
- 1080 Burchard, H., and R. Hofmeister (2008), A dynamic equation for the potential energy
1081 anomaly for analysing mixing and stratification in estuaries and coastal seas, *Estuarine,*
1082 *coastal and shelf science*, *77*(4), 679–687.

- 1083 Callies, J., & Ferrari, R. (2018). Baroclinic instability in the presence of convection.
1084 *Journal of Physical Oceanography*, *48*, 45–60.
- 1085 Capet, X., McWilliams, J. C., Molemaker, M. J., & Shchepetkin, A. (2008a). Mesoscale
1086 to submesoscale transition in the california current system. part ii: Frontal processes.
1087 *Journal of Physical Oceanography*, *38*, 44–64.
- 1088 Capet, X., McWilliams, J. C., Molemaker, M. J., & Shchepetkin, A. F. (2008b). Mesoscale
1089 to submesoscale transition in the california current system. part i: Flow structure, eddy
1090 flux, and observational tests. *J. Phys. Oceanogr.*, *38*, 29–43.
- 1091 Carranza, M. M., & Gille, S. T. (2015). Southern ocean wind-driven entrainment enhances
1092 satellite chlorophyll-a through the summer. *Journal of Geophysical Research: Oceans*,
1093 *120*, 304–323.
- 1094 Carranza, M. M., Gille, S. T., Franks, P. J. S., Johnson, K. S., Pinkel, R., & Girton, J. B.
1095 (2018). When mixed layers are not mixed. storm-driven mixing and bio-optical vertical
1096 gradients in mixed layers of the southern ocean. *Journal of Geophysical Research:*
1097 *Oceans*, *0*. doi:10.1029/2018JC014416.
- 1098 Chacko, N. (2017), Chlorophyll bloom in response to tropical cyclone hudhud in the bay
1099 of bengal: Bio-argo subsurface observations, *Deep Sea Research Part I: Oceanographic*
1100 *Research Papers*, *124*, 66–72.
- 1101 Crawford, G., & Large, W. (1996). A numerical investigation of resonant inertial response
1102 of the ocean to wind forcing. *J. Phys. Oceanogr.*, *26*, 873–891.
- 1103 Cushing, D. (1990). Plankton production and year-class strength in fish populations: an
1104 update of the match/mismatch hypothesis. *Advances in marine biology*, *26*, 249–293.

- 1105 D'Asaro, E., C. Lee, L. Rainville, R. Harcourt, and L. Thomas (2011), Enhanced turbu-
1106 lence and energy dissipation at ocean fronts, *Science*, *332*, 318–322.
- 1107 Doney, S. C., Glover, D. M., & Najjar, R. G. (1996). A new coupled, one-dimensional
1108 biological-physical model for the upper ocean: Applications to the jgofs bermuda at-
1109 lantic time-series study (bats) site. *Deep Sea Research Part II: Topical Studies in*
1110 *Oceanography*, *43*, 591–624.
- 1111 Erickson, Z. K., and A. F. Thompson (2018), The seasonality of physically driven export
1112 at submesoscales in the northeast atlantic ocean, *Global Biogeochemical Cycles*, *32*(8),
1113 1144–1162.
- 1114 Evans, G. T., & Garçon, V. (1997). One-dimensional models of water column biogeo-
1115 chemistry. *JGOFS report*, *23*, 85.
- 1116 Fasham, M., Ducklow, H., & McKelvie, S. (1990). A nitrogen-based model of plankton
1117 dynamics in the oceanic mixed layer. *Journal of Marine Research*, *48*, 591–639.
- 1118 Fasham, M., & Evans, G. (1995). The use of optimization techniques to model marine
1119 ecosystem dynamics at the jgofs station at 47 n 20 w. *Phil. Trans. R. Soc. Lond. B*,
1120 *348*, 203–209.
- 1121 Fauchereau, N., A. Tagliabue, L. Bopp, and P. M. Monteiro (2011), The response of
1122 phytoplankton biomass to transient mixing events in the southern ocean, *Geophysical*
1123 *Research Letters*, *38*(17).
- 1124 Fox-Kemper, B., Ferrari, R., & Hallberg, R. (2008). Parameterization of mixed layer
1125 eddies. part i: Theory and diagnosis. *J. Phys. Oceanogr.*, *38*, 1145–1165.
- 1126 Franks, P. J. (2002). Npz models of plankton dynamics: their construction, coupling to
1127 physics, and application. *Journal of Oceanography*, *58*, 379–387.

- 1128 Franks, P. J., & Walstad, L. J. (1997). Phytoplankton patches at fronts: a model of
1129 formation and response to wind events. *Journal of Marine Research*, *55*, 1–29.
- 1130 Girishkumar, M., V. Thangaprakash, T. Udaya Bhaskar, K. Suprit, N. Sureshkumar,
1131 S. Baliar Singh, J. Jofia, V. Pant, S. Vishnu, G. George, et al. (2019), Quantifying trop-
1132 ical cyclone’s effect on the biogeochemical processes using profiling float observations
1133 in the bay of bengal, *Journal of Geophysical Research: Oceans*.
- 1134 Hamlington, P. E., VanRoekel, L. P., Fox-Kemper, B., Julien, K., & Chini, G. (2014).
1135 Langmuir-submesoscale interactions: Descriptive analysis of multiscale frontal spindown
1136 simulations. *J. Phys. Oceanogr.*, *44*, 2249–2272.
- 1137 Haney, S., Fox-Kemper, B., Julien, K., & Webb, A. (2015). Symmetric and geostrophic
1138 instabilities in the wave-forced ocean mixed layer. *Journal of Physical Oceanography*,
1139 *45*, 3033–3056.
- 1140 Kaltenbach, H.-J., Gerz, T., & Schumann, U. (1994). Large-eddy simulation of homo-
1141 geneous turbulence and diffusion in stably stratified shear flow. *J. Fluid Mech.*, *280*,
1142 1–40.
- 1143 Klein, P., & Lapeyre, G. (2009). The oceanic vertical pump induced by mesoscale and
1144 submesoscale turbulence. *Annual Review of Marine Science*, *1*, 351–375.
- 1145 Kraus, E. B., and J. S. Turner (1967), A one-dimensional model of the seasonal thermo-
1146 cline ii. the general theory and its consequences, *Tellus*, *19*(1), 98–106.
- 1147 Large, W., & Crawford, G. (1995). Observations and simulations of upper-ocean response
1148 to wind events during the ocean storms experiment. *Journal of Physical Oceanography*,
1149 *25*, 2831–2852.

- 1150 Large, W. G., McWilliams, J. C., & Doney, S. C. (1994). Oceanic vertical mixing: A
1151 review and a model with a nonlocal boundary layer parameterization. *Rev. Geophys.*,
1152 *32*, 363–403.
- 1153 Lévy, M., Ferrari, R., Franks, P. J., Martin, A. P., & Rivière, P. (2012). Bringing physics
1154 to life at the submesoscale. *Geophys. Res. Lett.*, *39*.
- 1155 Lévy, M., Franks, P. J., & Smith, K. S. (2018). The role of submesoscale currents in
1156 structuring marine ecosystems. *Nature communications*, *9*, 4758.
- 1157 Lévy, M., Klein, P., & Ben Jelloul, M. (2009). New production stimulated by high-
1158 frequency winds in a turbulent mesoscale eddy field. *Geophysical Research Letters*,
1159 *36*.
- 1160 Lévy, M., Klein, P., & Treguier, A.-M. (2001). Impact of sub-mesoscale physics on pro-
1161 duction and subduction of phytoplankton in an oligotrophic regime. *J. Mar. Res.*, *59*,
1162 535–565.
- 1163 Lévy, M., & Martin, A. (2013). The influence of mesoscale and submesoscale heterogeneity
1164 on ocean biogeochemical reactions. *Global Biogeochemical Cycles*, *27*, 1139–1150.
- 1165 Lewis, D., Brereton, A., & Siddons, J. (2017). A large eddy simulation study of the
1166 formation of deep chlorophyll/biological maxima in un-stratified mixed layers: The
1167 roles of turbulent mixing and predation pressure. *Limnology and Oceanography*, *62*,
1168 2277–2307.
- 1169 Li, K., Zhang, Z., Chini, G., & Flierl, G. (2012). Langmuir circulation: an agent for
1170 vertical restratification? *Journal of Physical Oceanography*, *42*, 1945–1958.
- 1171 Lin, I.-I. (2012), Typhoon-induced phytoplankton blooms and primary productivity in-
1172 crease in the western north pacific subtropical ocean, *Journal of Geophysical Research*:

1173 *Oceans*, 117(C3).

1174 Lin, I., Lin, I., Liu, W., Wu, C., Wong, G., Hu, C., Chen, Z., Liang, W., Yang, Y., & Liu,
1175 K. (2003). New evidence for enhanced ocean primary production triggered by tropical
1176 cyclone *Geophysical Research Letters*, 30, 1718.

1177 Long, M. C., Thomas, L. N., & Dunbar, R. B. (2012). Control of phytoplankton bloom
1178 inception in the ross sea, antarctica, by ekman restratification. *Global Biogeochemical*
1179 *Cycles*, 26.

1180 Mahadevan, A. (2016). The impact of submesoscale physics on primary productivity of
1181 plankton. *Annual review of marine science*, 8, 161–184.

1182 Mahadevan, A., & Archer, D. (2000). Modeling the impact of fronts and mesoscale
1183 circulation on the nutrient supply and biogeochemistry of the upper ocean. *Journal of*
1184 *Geophysical Research: Oceans*, 105, 1209–1225.

1185 Mahadevan, A., D’Asaro, E., Lee, C., & Perry, M. J. (2012). Eddy-driven stratification
1186 initiates north atlantic spring phytoplankton blooms. *Science*, 337, 54–58.

1187 Mahadevan, A., & Tandon, A. (2006). An analysis of mechanisms for submesoscale vertical
1188 motion at ocean fronts. *Ocean Modelling*, 14, 241–256.

1189 Mahadevan, A., Tandon, A., & Ferrari, R. (2010). Rapid changes in mixed layer stratifi-
1190 cation driven by submesoscale instabilities and winds. *J. Geophys. Res.*, 115.

1191 Mahadevan, A., Thomas, L. N., & Tandon, A. (2008). Comment on “Eddy/Wind inter-
1192 actions stimulate extraordinary mid-ocean plankton blooms”. *Science*, 320, 448–448.

1193 Marra, J., Bidigare, R., & Dickey, T. (1990). Nutrients and mixing, chlorophyll and
1194 phytoplankton growth. *Deep Sea Research*, 37, 127–143.

- 1195 Martin, A. P., Lévy, M., Gennip, S., Pardo, S., Srokosz, M., Allen, J., Painter, S. C.,
1196 & Pidcock, R. (2015). An observational assessment of the influence of mesoscale and
1197 submesoscale heterogeneity on ocean biogeochemical reactions. *Global Biogeochemical*
1198 *Cycles*, *29*, 1421–1438.
- 1199 McGillicuddy, D., McCarthy, J., & Robinson, A. (1995). Coupled physical and biological
1200 modeling of the spring bloom in the north atlantic (i): model formulation and one di-
1201 mensional bloom processes. *Deep Sea Research Part I: Oceanographic Research Papers*,
1202 *42*, 1313–1357.
- 1203 McWilliams, J. C., & Sullivan, P. P. (2000). Vertical mixing by langmuir circulations.
1204 *Spill Science & Technology Bulletin*, *6*, 225–237.
- 1205 Nagai, T., Clayton, S. (2017). Nutrient interleaving below the mixed layer of the Kuroshio
1206 Extension Front *Ocean Dynamics*, *67*, 1027–1046.
- 1207 Nicholson, S.-A., Lévy, M., Llort, J., Swart, S., & Monteiro, P. M. S. (2016). In-
1208 vestigation into the impact of storms on sustaining summer primary productivity in
1209 the sub-antarctic ocean. *Geophysical Research Letters*, *43*, 9192–9199. doi:10.1002/
1210 2016GL069973.
- 1211 Ocean Ecology Laboratory, Ocean Biology Processing Group NASA Goddard Space Flight
1212 Center. (2018). Visible and infrared imager/radiometer suite (viirs) photosynthetically
1213 available radiation data; 2018 reprocessing. doi:10.5067/NPP/VIIRS/L3B/PAR/2018
1214 nASA OB.DAAC, Greenbelt, MD, USA. Accessed on 03/16/2018.
- 1215 Painter, S. C., Finlay, M., Hemsley, V. S., & Martin, A. P. (2016). Seasonality, phytoplank-
1216 ton succession and the biogeochemical impacts of an autumn storm in the northeast
1217 atlantic ocean. *Progress in Oceanography*, *142*, 72–104.

- 1218 Pham, H., & Sarkar, S. (2017). Turbulent entrainment in a strongly stratified barrier
1219 layer. *Journal of Geophysical Research: Oceans*, *122*, 5075–5087.
- 1220 Platt, T., Fuentes-Yaco, C., & Frank, K. T. (2003). Marine ecology: spring algal bloom
1221 and larval fish survival. *Nature*, *423*, 398–399.
- 1222 Pollard, R. T., P. B. Rhines, and R. Thompson (1972), The deepening of the wind-mixed
1223 layer, *Geophysical & Astrophysical Fluid Dynamics*, *4*(1), 381–404.
- 1224 Powell, T. M., Lewis, C. V., Curchitser, E. N., Haidvogel, D. B., Hermann, A. J., &
1225 Dobbins, E. L. (2006). Results from a three-dimensional, nested biological-physical
1226 model of the california current system and comparisons with statistics from satellite
1227 imagery. *Journal of Geophysical Research: Oceans*, *111*.
- 1228 Rumyantseva, A., Lucas, N., Rippeth, T., Martin, A., Painter, S. C., Boyd, T. J., &
1229 Henson, S. (2015). Ocean nutrient pathways associated with the passage of a storm.
1230 *Global Biogeochemical Cycles*, *29*, 1179–1189.
- 1231 Skillingstad, E., & Samelson, R. (2012). Baroclinic frontal instabilities and turbulent
1232 mixing in the surface boundary layer. part i: unforced simulations. *J. Phys. Oceanogr.*,
1233 *42*, 1701–1715.
- 1234 Skillingstad, E. D., Duncombe, J., & Samelson, R. M. (2017). Baroclinic frontal insta-
1235 bilities and turbulent mixing in the surface boundary layer. part ii: Forced simulations.
1236 *Journal of Physical Oceanography*, *47*, 2429–2454.
- 1237 Skillingstad, E. D., Smyth, W., & Crawford, G. (2000). Resonant wind-driven mixing in
1238 the ocean boundary layer. *Journal of physical oceanography*, *30*, 1866–1890.
- 1239 Smith, K. M., Hamlington, P. E., & Fox-Kemper, B. (2016). Effects of submesoscale tur-
1240 bulence on ocean tracers. *J. Geophys. Res.*, *121*, 908–933. doi:10.1002/2016JC012400.

- 1241 Stone, P. H. (1966). On non-geostrophic baroclinic stability. *J. Atmos. Sci.*, *23*, 390–400.
- 1242 Taylor, J. (2016). Turbulent mixing, restratification, and phytoplankton growth at a
1243 submesoscale eddy. *Geophys. Res. Lett.*, .
- 1244 Taylor, J. R. (2008). *Numerical simulations of the stratified oceanic bottom boundary*
1245 *layer*. University of California, San Diego.
- 1246 Taylor, J. R., & Ferrari, R. (2010). Buoyancy and wind-driven convection at mixed layer
1247 density fronts. *J. Phys. Oceanogr.*, *40*, 1222–1242.
- 1248 Thomalla, S., Fauchereau, N., Swart, S., & Monteiro, P. (2011). Regional scale charac-
1249 teristics of the seasonal cycle of chlorophyll in the southern ocean. *Biogeosciences*, *8*,
1250 2849.
- 1251 Thomas, L., & Ferrari, R. (2008). Friction, frontogenesis, and the stratification of the
1252 surface mixed layer. *J. Phys. Oceanogr.*, *38*, 2501–2518.
- 1253 Thomas, L., Tandon, A., & Mahadevan, A. (2008). Submesoscale processes and dynamics.
1254 *Ocean Modeling in an Eddy Regime*, *177*, 17–37.
- 1255 Thomas, L., & Taylor, J. R. (2010). Reduction of the usable wind-work on the general
1256 circulation by forced symmetric instability. *Geophys. Res. Lett.*, *37*.
- 1257 Thomas, L. N., J. R. Taylor, E. A. D’Asaro, C. M. Lee, J. M. Klymak, and A. Shcherbina
1258 (2016), Symmetric instability, inertial oscillations, and turbulence at the Gulf Stream
1259 front, *J. Phys. Oceanogr.*, (2016), *46*, 197–217.
- 1260 Thompson, A. F., Lazar, A., Buckingham, C., Naveira-Garabato, A. C., Damerell, G. M.,
1261 & Heywood, K. J. (2016). Open-ocean submesoscale motions: A full seasonal cycle of
1262 mixed layer instabilities from gliders. *J. Phys. Oceanogr.*, *46*, 1285–1307.

- 1263 Waniek, J. J. (2003). The role of physical forcing in initiation of spring blooms in the
1264 northeast atlantic. *Journal of Marine Systems*, *39*, 57–82.
- 1265 Whitt, D. (2017). A one-dimensional npzd model in matlab, v1.0. doi:doi:10.5281/
1266 zenodo.569312.
- 1267 Whitt, D. (2018). Data for “Energetic submesoscales maintain strong mixed layer strati-
1268 fication during an autumn storm”. doi:10.6084/m9.figshare.c.3816175.v3 accessed
1269 December 6, 2018.
- 1270 Whitt, D., Lévy, M., & Taylor, J. (2017a). Low-frequency and high-frequency oscilla-
1271 tory winds synergistically enhance nutrient entrainment and phytoplankton at fronts.
1272 *Journal of Geophysical Research: Oceans*, *122*, 1016–1041.
- 1273 Whitt, D. B., & Taylor, J. R. (2017). Energetic submesoscales maintain strong mixed
1274 layer stratification during an autumn storm. *Journal of Physical Oceanography*, *47*,
1275 2419–2427.
- 1276 Whitt, D. B., Taylor, J. R., & Lévy, M. (2017b). Synoptic-to-planetary scale wind
1277 variability enhances phytoplankton biomass at ocean fronts. *J. Geophys. Res.*, *122*.
1278 doi:doi:10.1002/2016JC011899.
- 1279 Woodson, C. B., & Litvin, S. Y. (2015). Ocean fronts drive marine fishery production
1280 and biogeochemical cycling. *Proceedings of the National Academy of Sciences*, *112*,
1281 1710–1715.
- 1282 Wyngaard, J. C. (2004), Toward numerical modeling in the terra incognita; *Journal of*
1283 *the atmospheric sciences*, *61*(14), 1816–1826.
- 1284 Ye, H., Y. Sui, D. Tang, and Y. Afanasyev (2013), A subsurface chlorophyll a bloom
1285 induced by typhoon in the south china sea, *Journal of Marine Systems*, *128*, 138–145.

Table 1. List of simulations

Simulation and parameterizations	grid points (x,y,z)	grid resolution	max $ \tau $	$\langle M^2 \rangle_{x,y}$
1. LES with front (LES,F)	1024×1024×160	1.9×1.9×0.5 m	0.3 N/m ²	$5.0 \times 10^{-8} \text{ s}^{-2}$
2. Column model with front, KPP, and FFH (ROMS,F)	2×2×160	39×39×0.5 m	0.3 N/m ²	$5.0 \times 10^{-8} \text{ s}^{-2}$
3. LES without front (LES,NF)	256×256×160	1.9×1.9×0.5 m	0.3 N/m ²	0 s ⁻²
4. Column model with KPP (ROMS,NF)	2×2×160	39×39×0.5 m	0.3 N/m ²	0 s ⁻²
5. LES with front (LES,F)	1024×1024×160	1.9×1.9×0.5 m	0.6 N/m ²	$5.0 \times 10^{-8} \text{ s}^{-2}$
6. Column model with front, KPP, and FFH (ROMS,F)	2×2×160	39×39×0.5 m	0.6 N/m ²	$5.0 \times 10^{-8} \text{ s}^{-2}$
7. LES without front (LES,NF)	256×256×160	1.9×1.9×0.5 m	0.6 N/m ²	0 s ⁻²
8. Column model with KPP (ROMS,NF)	2×2×160	39×39×0.5 m	0.6 N/m ²	0 s ⁻²

Table 2. Biogeochemical Model Parameters

Parameter	Symbol	Value
Light attenuation due to seawater	k_w	0.055 m ⁻¹
Initial slope of the productivity-irradiance curve	α	0.1 d ⁻¹ (W/m ²) ⁻¹
Surface photosynthetically available radiation	I_0	50 W/m ²
Phytoplankton max. uptake rate	V_m	1.5 d ⁻¹
Half-saturation for nutrient uptake by phy.	k_N	0.8 mmol /m ³
Phytoplankton mortality rate to detritus	σ_d	0.04 d ⁻¹
Ivlev constant for zooplankton grazing	Λ	0.5 (mmol /m ³) ⁻¹
Maximum zooplankton grazing rate	R	0.9 d ⁻¹
Zooplankton excretion to nutrient	γ_N	0.3
Linear zooplankton mortality rate	ζ	0.04 d ⁻¹
Quadratic zooplankton mortality rate	$\hat{\zeta}$	0.28 d ⁻¹ (mmol/m ³) ⁻¹
Detritus remineralization rate	δ	0.1 d ⁻¹
Detrital sinking velocity	w_d	-0.25 m/d
Restoring nutrient concentration at 80 m	N_{80}	2.5 mmol/m ³
Inverse restoring timescale for nutrient at 80 m	β_N	0.25 d ⁻¹

Figure 1. Slices from showing the snapshots of the density and vertical velocity in a large eddy simulation of the frontal zone, just before the storm forcing begins. The horizontal slices (top panels) are at 21 m depth and the vertical slices (bottom panels) are 947 m across the front, as indicated by the dashed lines. Density anomaly (a) and vertical velocity (b) are shown. Black contours of the density (smoothed by convolving with a 150 m square boxcar kernel) are plotted every 0.0025 kg/m³ on all panels. However, only the least dense 0.035 kg/m³ of the density range is contoured in order to highlight the density variability in the mixed layer rather than the pycnocline. The bottom panels show only the top 50 m to highlight the surface properties of interest.

1286

Figure 2. Time series of the magnitude of the surface stress (solid lines), which is directed to the southeast 45° to the right of the mean geostrophic flow at the surface in both scenarios. The atmospheric and Ekman buoyancy flux B_A+EBF , where $B_A = 3 \times 10^{-9} \text{ m}^2/\text{s}^3$ is constant, is indicated by the dotted lines. The thick black lines indicate two different averaging windows used in the subsequent analysis: days 2-3 and days 5-6.

Figure 3. Solid lines are initial model profiles of phytoplankton P (mmol/m^3), nutrient N (mmol/m^3), and density ρ (kg/m^3). The model currency for N and P is the same: mmol/m^3 of a single generalized “nutrient.” Solid and dotted black horizontal lines indicate the initial depths of MLD_{05} and MLD_3 , respectively. Scattered black points are observed in-situ chlorophyll-a fluorescence (mg/m^3) (a), dissolved inorganic silica (mmol/m^3) (b), and the potential density (kg/m^3) (c) from 3 profiles taken above the Porcupine Abyssal Plain during September 21-22, 2012 on cruise D381B [Painter et al., 2016]. Biogenic silica particulate concentrations (not shown) are scattered between about 0.05 and 0.25 mmol/m^3 , and exhibit no obvious subsurface maximum like chlorophyll-a fluorescence [Fig. 14a in Painter et al., 2016]. Only the top 60 m is shown to highlight the surface properties of interest.

Figure 4. Horizontally-averaged vertical profiles of phytoplankton $\langle P \rangle_{x,y}$, nutrient $\langle N \rangle_{x,y}$, and density $\langle \rho \rangle_{x,y}$ averaged over two time periods during the evolution of the idealized storm life cycle, including $t = 2 - 3$ days (during the storm), and $t = 5 - 6$ days (after the storm). Mixed layer depths based on the $\Delta \langle \rho \rangle_{x,y} = 0.005 \text{ kg/m}^3$ threshold (MLD_{05} , thin solid) and $\Delta \langle \rho \rangle_{x,y} = 0.03 \text{ kg/m}^3$ threshold (MLD_3 , dotted) are also shown. Note the dotted lines essentially overlap, that is MLD_3 is about the same in both the F and NF cases.

Figure 5. (a) Time-integrated wind work, $\Delta KE = \frac{1}{1 \text{ day}} \int_{t=5 \text{ days}}^{t=6 \text{ days}} \int_0^t \boldsymbol{\tau} \cdot \mathbf{u}_h(z=0) ds dt$ where s is a dummy time integration variable, and (b) the mixing ratio $r_m = \Delta PE^* / \Delta KE$ where ΔPE^* is the residual change in potential energy due to vertical mixing, which is defined in (4). In both ΔKE and ΔPE^* , Δ indicates that the difference is taken between an average over $t = 5 - 6$ days and $t = 0$ days.

Figure 6. Full-depth column integrals of phytoplankton $\langle P \rangle_{x,y}$ (a) and storm-driven $\langle NCP \rangle_{x,y}$ (b), averaged over three time periods. The reduction in column-integrated $\langle N \rangle_{x,y}$ is equal to the storm-driven $\langle NCP \rangle_{x,y}$ here.

Figure 7. Slices of nutrient concentration N (a),(c) and vertical nutrient flux wN (b),(d) during the storm forcing from the LES,F (a)-(b) and LES,NF (c)-(d) scenarios. The horizontal slices (top panels) are at -39.5 m and -38.5 m respectively, that is 4 m above MLD_3 in both cases. The horizontal slices are at the depths indicated by the black dashed lines in the adjacent vertical slices/bottom panels, while the vertical slices are at the location indicated by the dashed lines in the adjacent horizontal slices. Solid black contours of density (smoothed with a 150 m moving average) are plotted every 0.005 kg/m^3 on all panels. The vertical slices show only the top 50 m to highlight the surface properties of interest. The vertical velocities (not shown) are an order of magnitude stronger than in Fig. 1 (b).

Figure 8. Multi-scale vertical profiles of horizontal nutrient variance $var(N)_{x,y}(z)$, both during (a),(c),(e) and after (b),(d),(f) the storm. Submesoscales (a)-(b), that is wavelengths between 0.15 and 2 km, are separated from smaller-turbulent scales (c)-(d), that is wavelengths < 0.15 km. The mixed layer depths MLD_3 (dotted) and MLD_{05} (solid) are plotted as thin horizontal lines.

Figure 9. Multi-scale vertical nutrient fluxes $\langle wN \rangle_{x,y}(z)$, both during (a),(c),(e) and after (b),(d),(f) the storm. Submesoscale fluxes are shown in (a)-(b), turbulent scale fluxes are shown in (c)-(d), and total (submesoscale+turbulent) are shown in (e)-(f). The mixed layer depths MLD_3 (dotted) and MLD_{05} (solid) are plotted as thin horizontal lines. Note the different x -axes.

Figure 10. Horizontally-averaged profiles of the total vertical flux (a)-(b) and diffusivity (c)-(d) of nutrient in the LES,F and ROMS,F scenarios with a front/submesoscales. MLD_3 (dotted) and MLD_{05} (solid) are plotted for reference. Note the different x-axes in (a)-(b).

Figure 11. Horizontally-averaged profiles of the total vertical flux (a)-(b) and diffusivity (c)-(d) of nutrient in the LES,NF and ROMS,NF scenarios without a front/submesoscales. MLD_3 (dotted) and MLD_{05} (solid) are plotted for reference. Note the different x-axes in (a)-(b).

Figure 12. Comparisons between LES and ROMS with (a)-(b) and without (c)-(d) a front/submesoscales. Full-depth column integrals of $\langle P \rangle_{x,y}$ (a),(c) and $\langle NCP \rangle_{x,y}$ (b),(d). All variables are presented as averages over three different time windows, as in Fig. 6.

Figure 1.

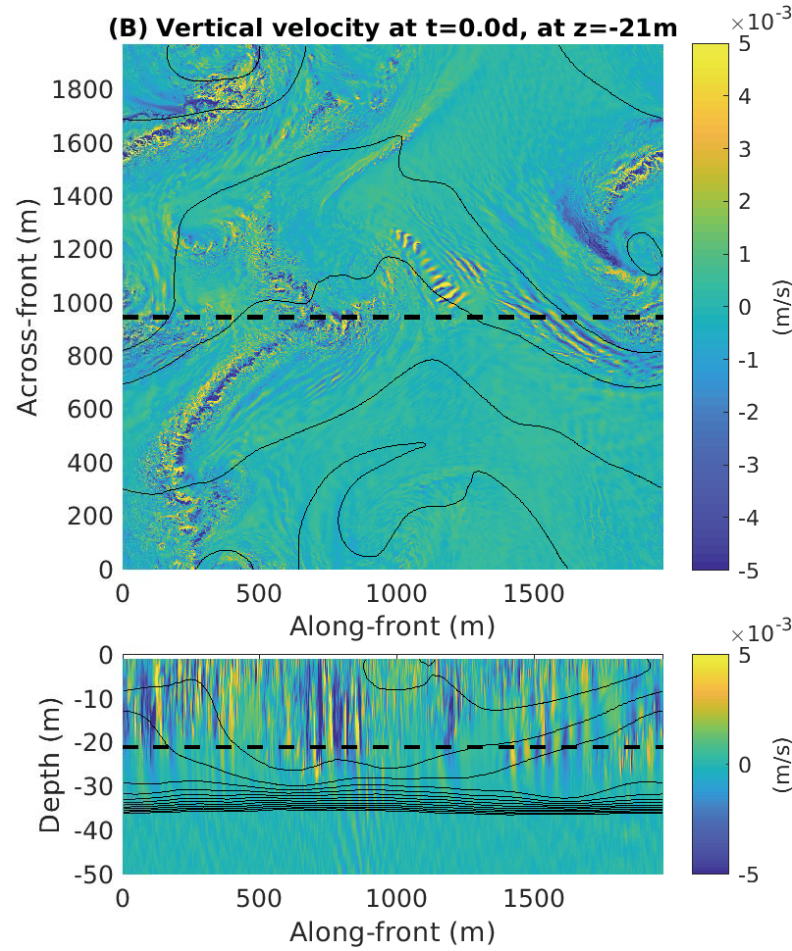
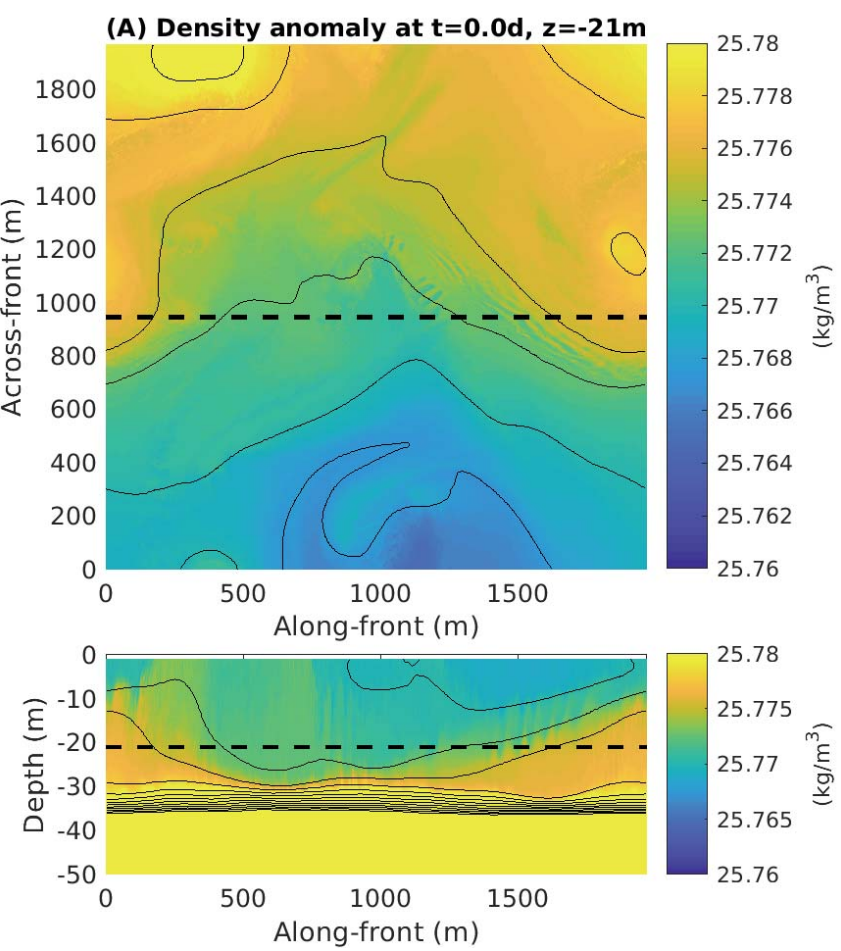


Figure 2.

Surface stress and Atmospheric+Ekman buoyancy flux

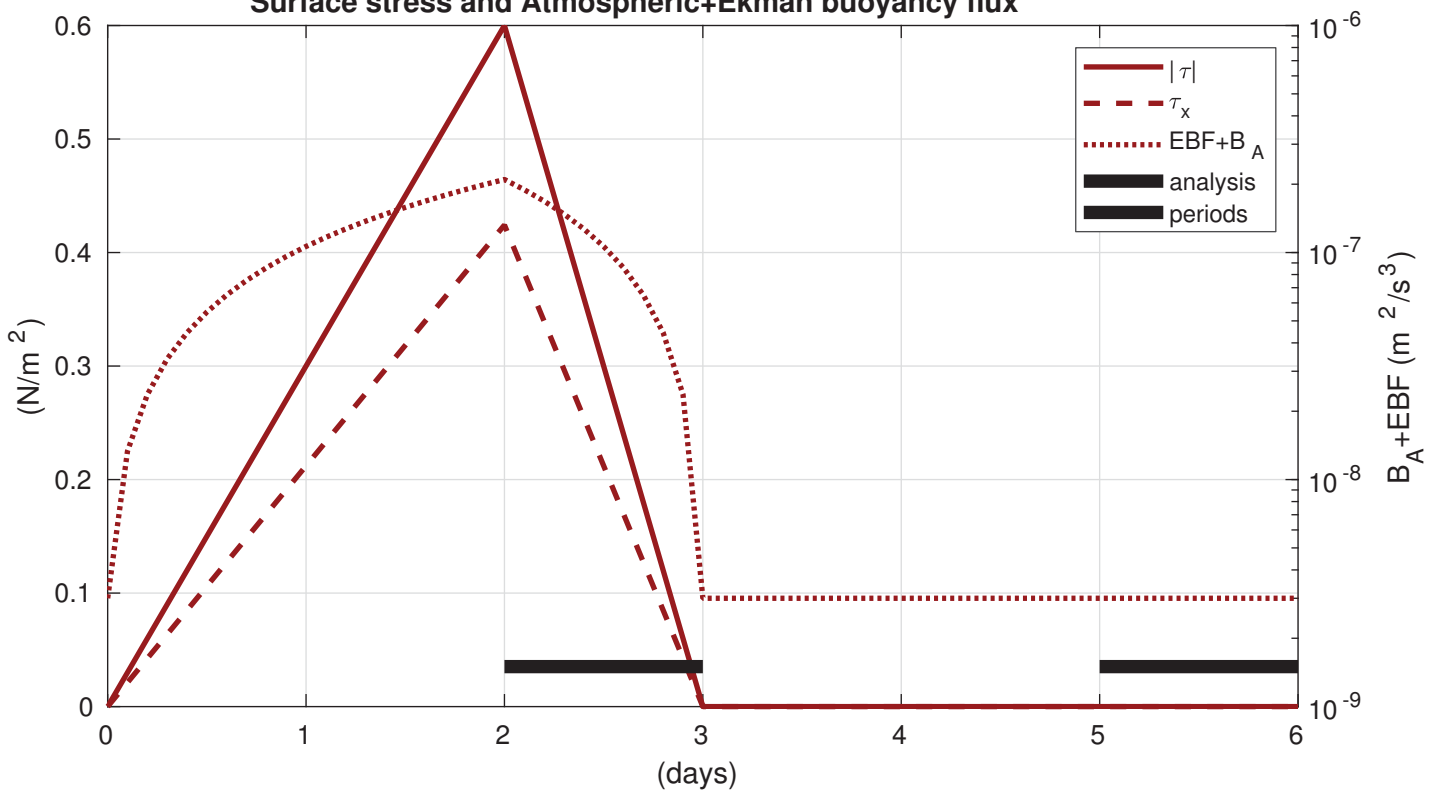
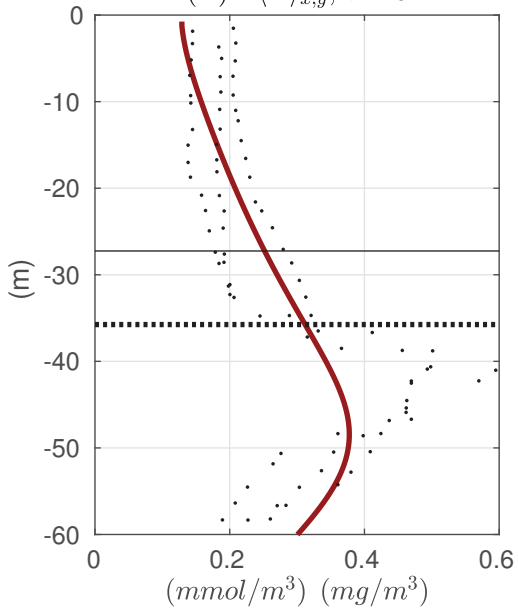
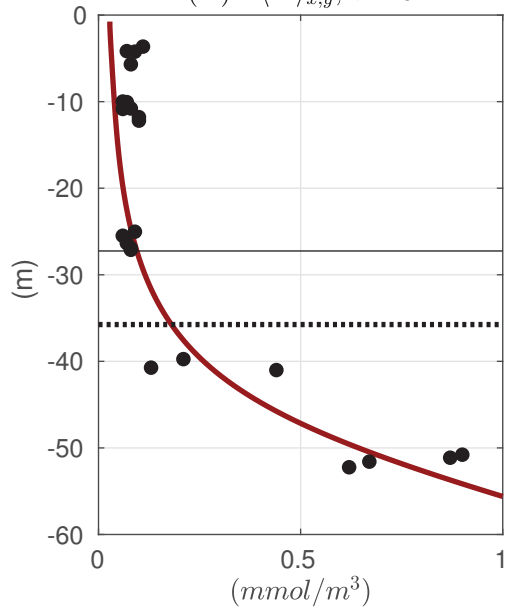


Figure 3.

(A) $\langle P \rangle_{x,y}, t = 0$



(B) $\langle N \rangle_{x,y}, t = 0$



(C) $\langle \rho \rangle_{x,y}, t = 0$

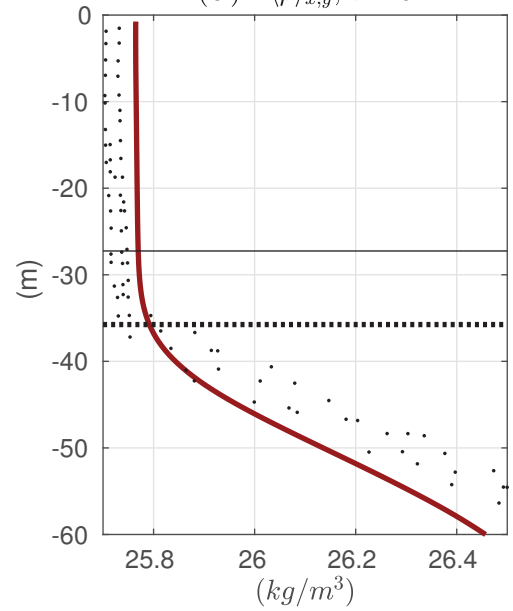


Figure 4.

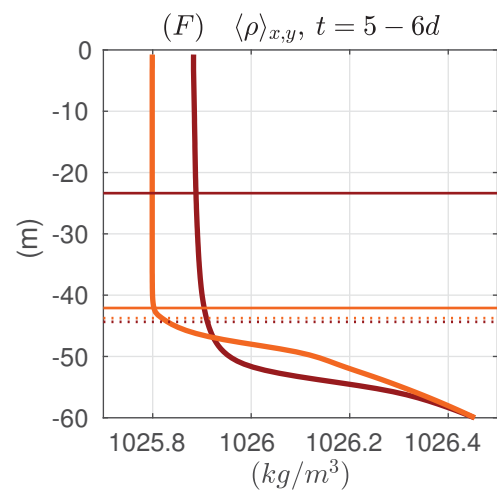
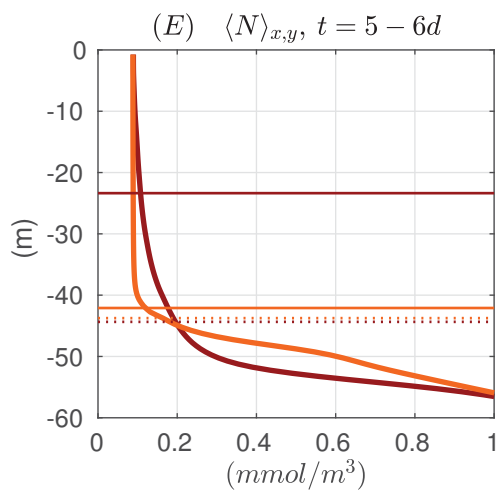
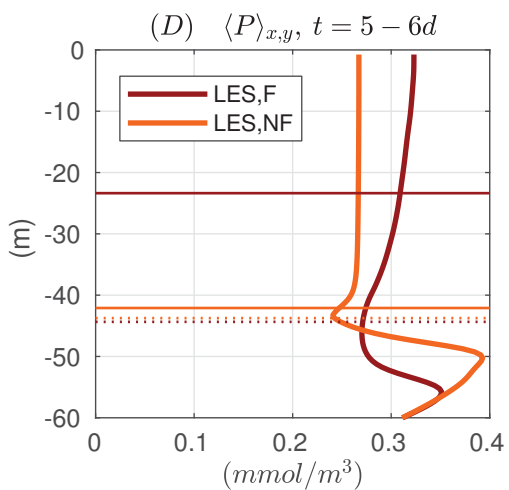
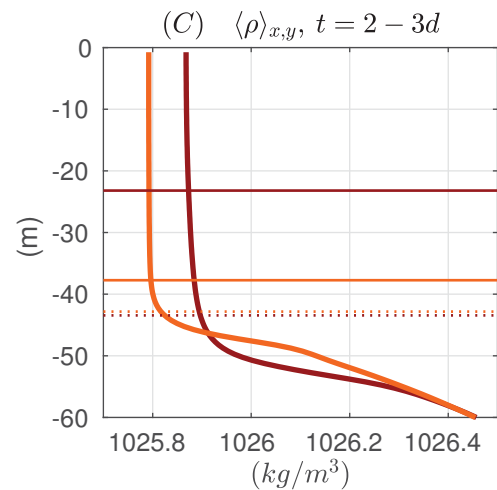
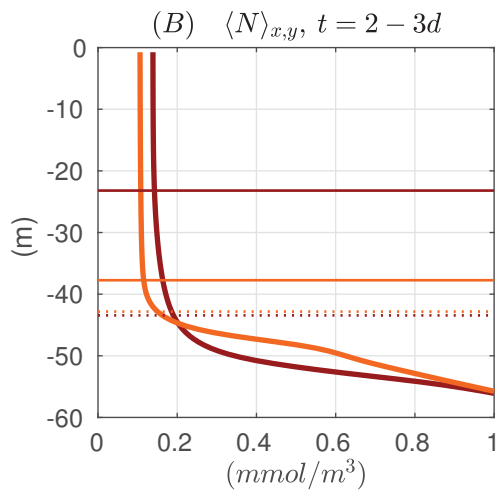
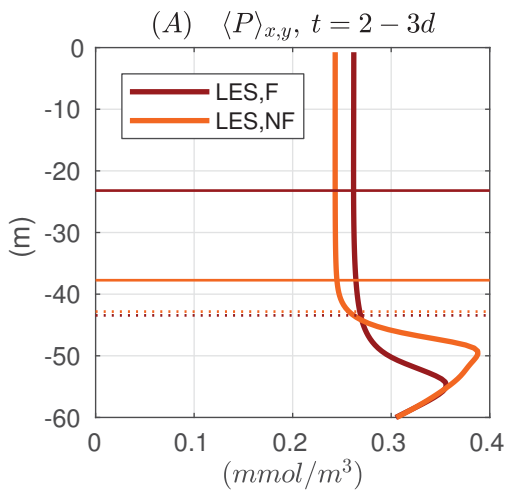


Figure 5.

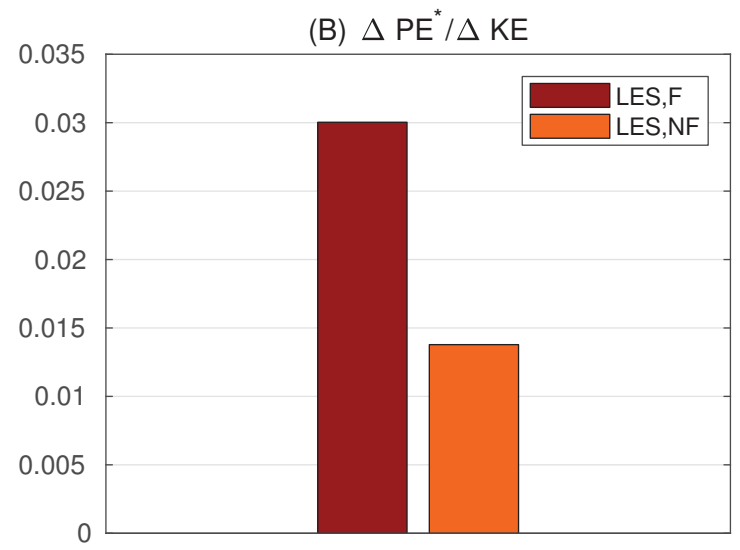
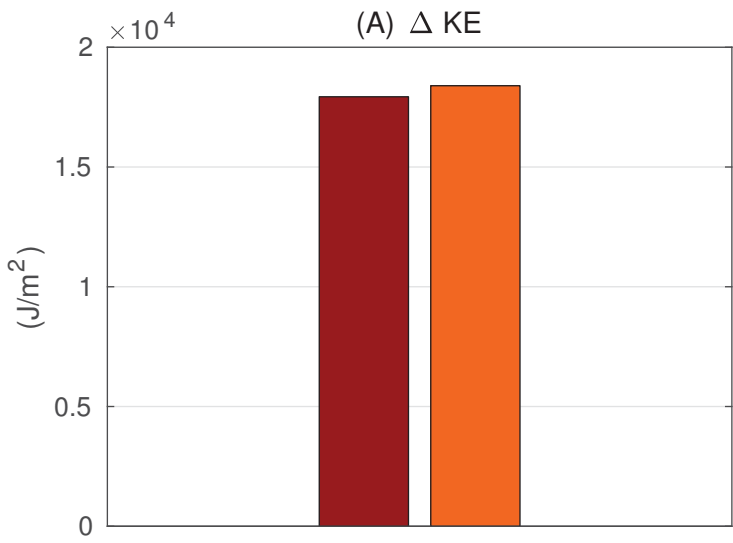


Figure 6.

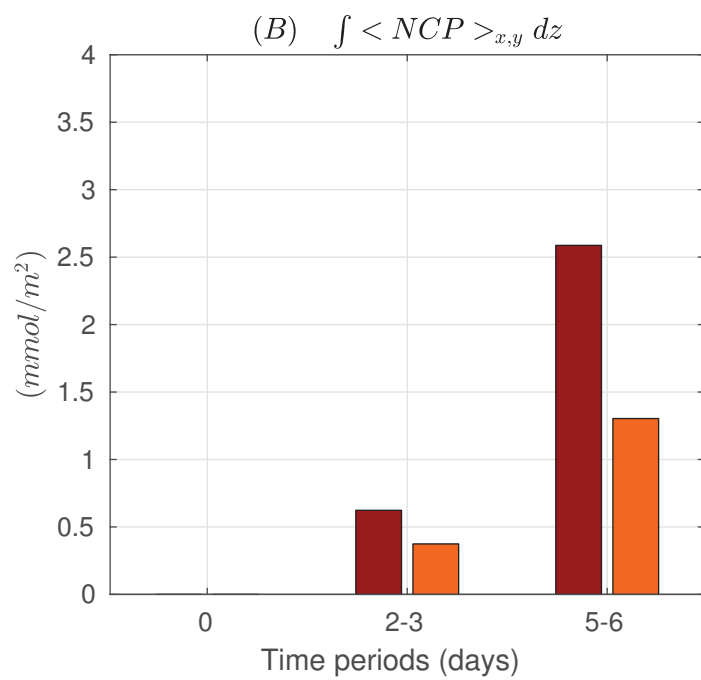
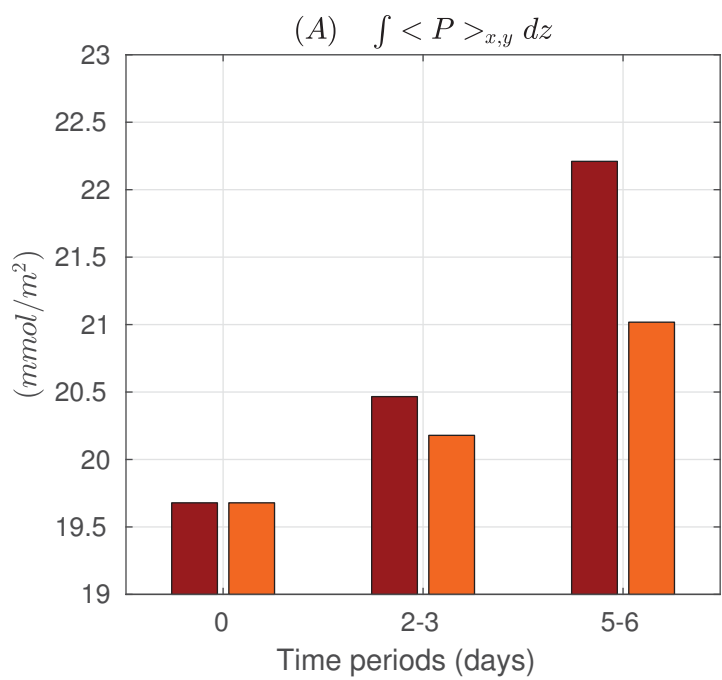


Figure 7.

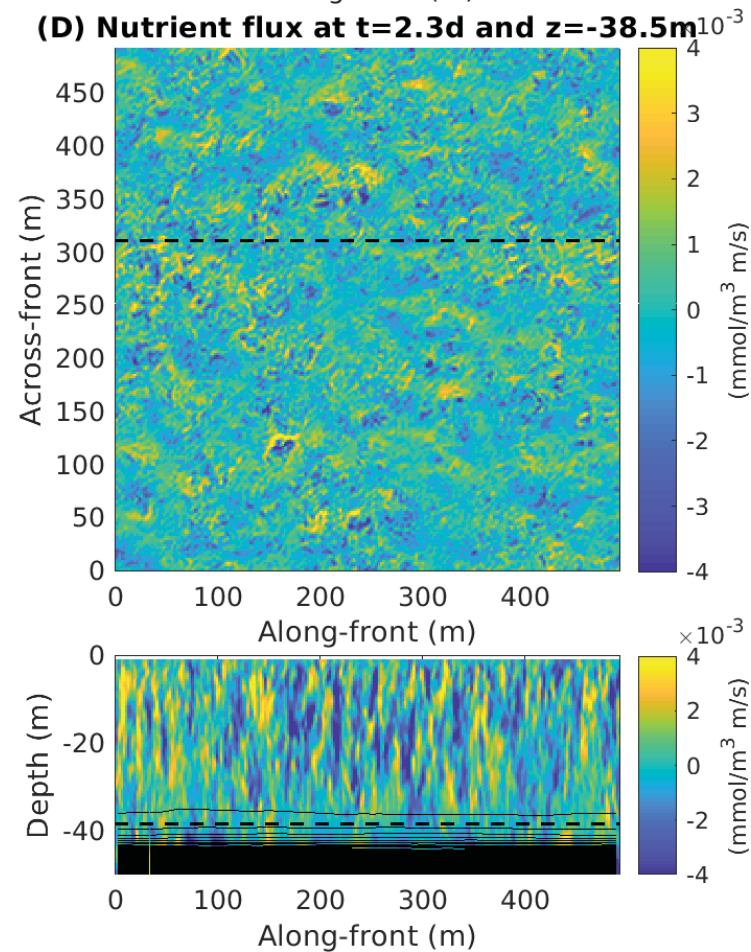
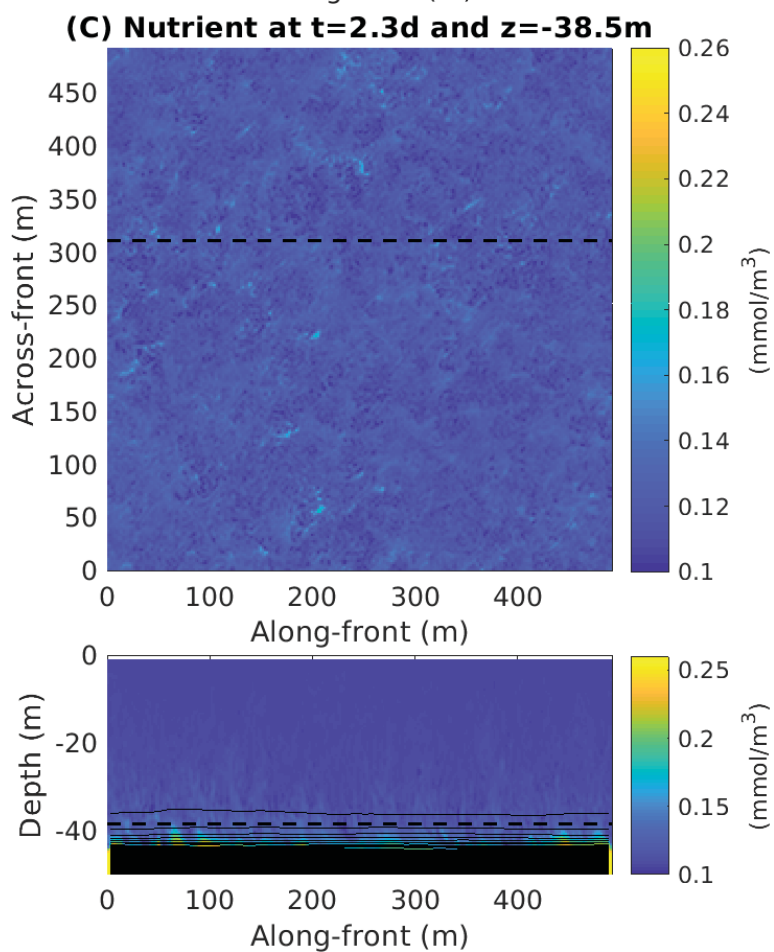
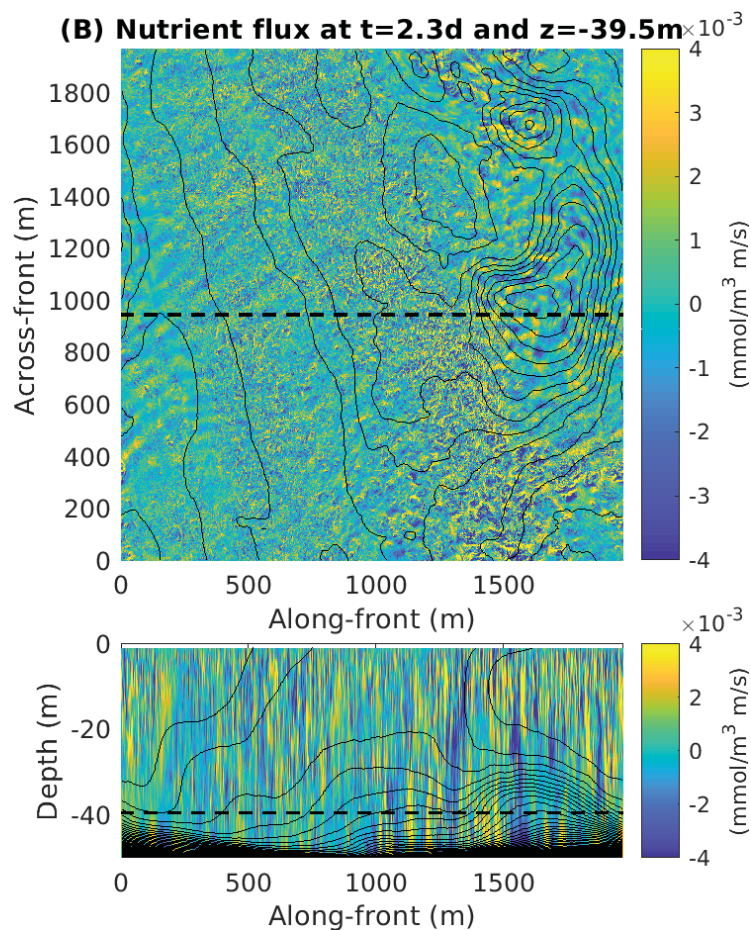
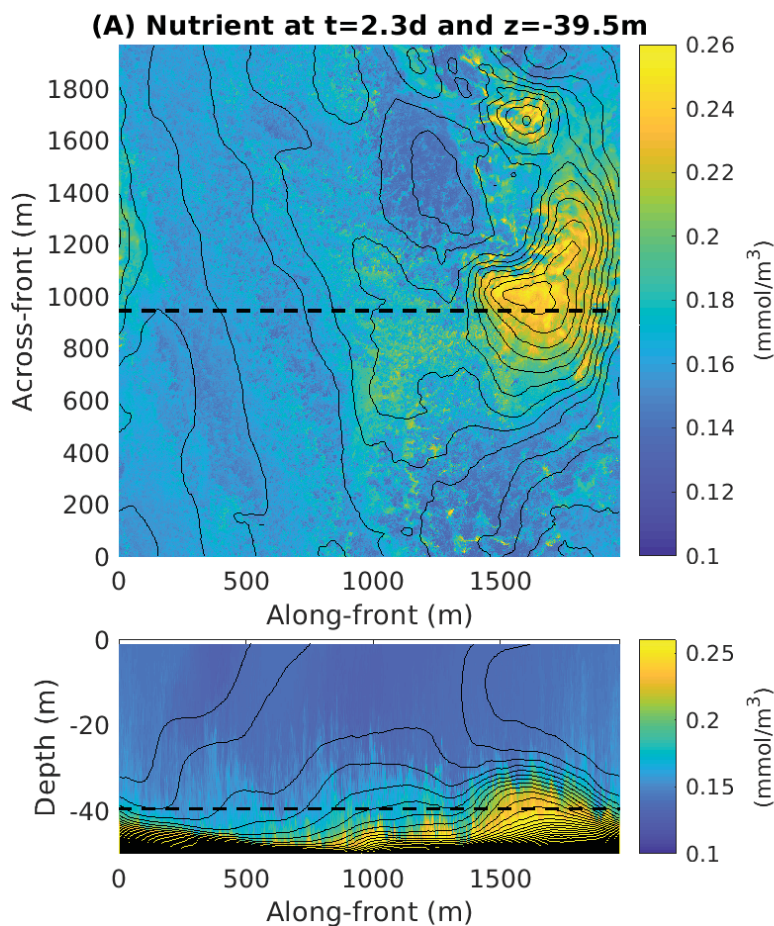


Figure 8.

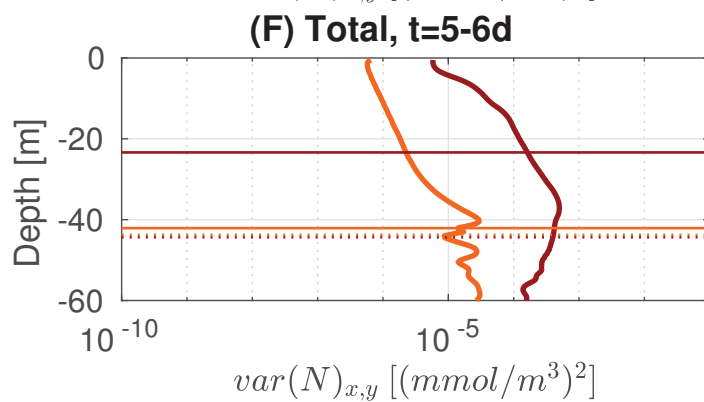
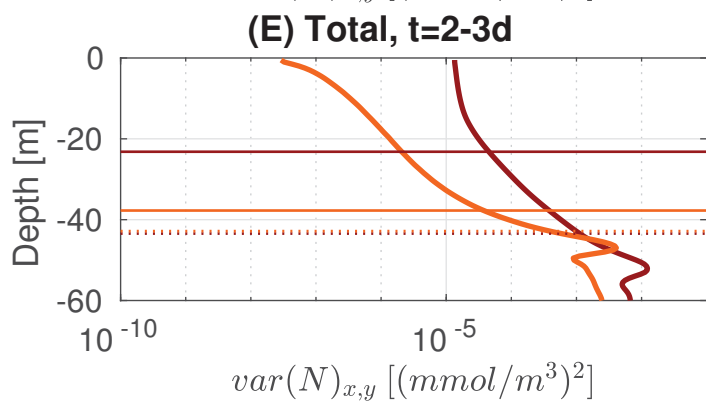
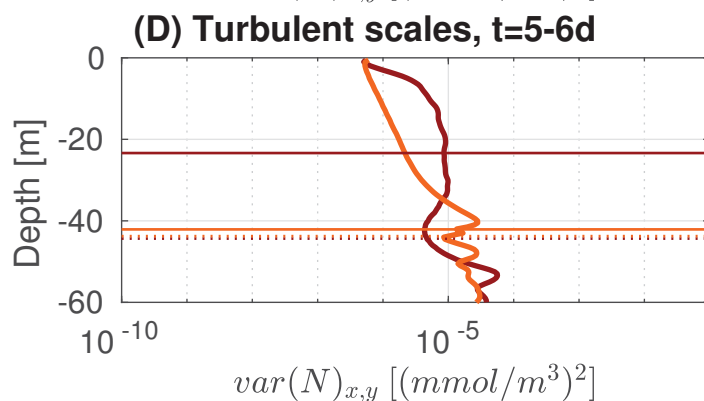
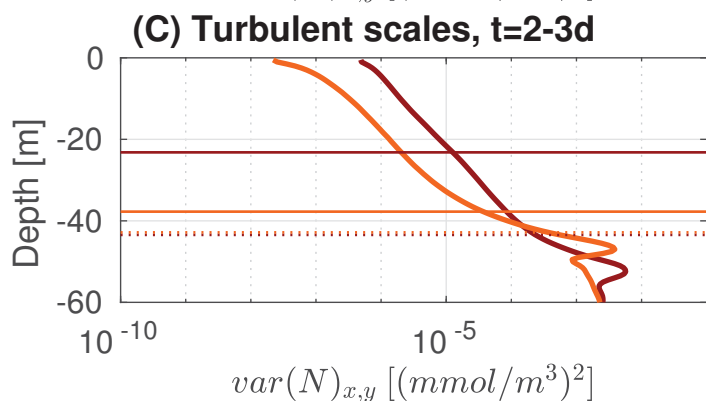
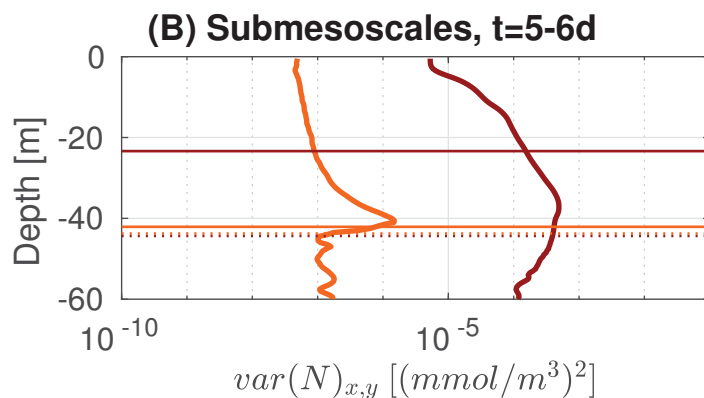
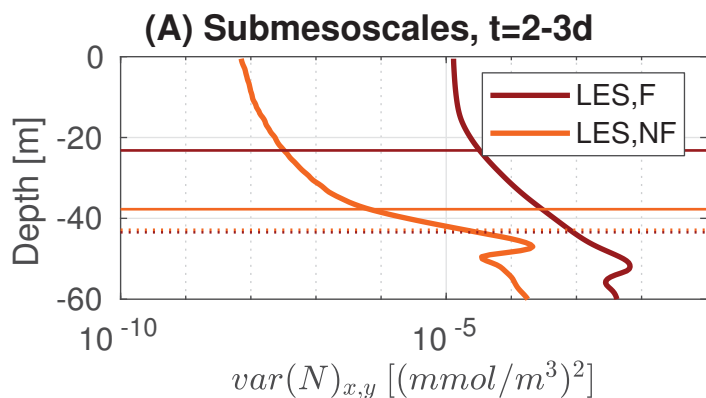


Figure 9.

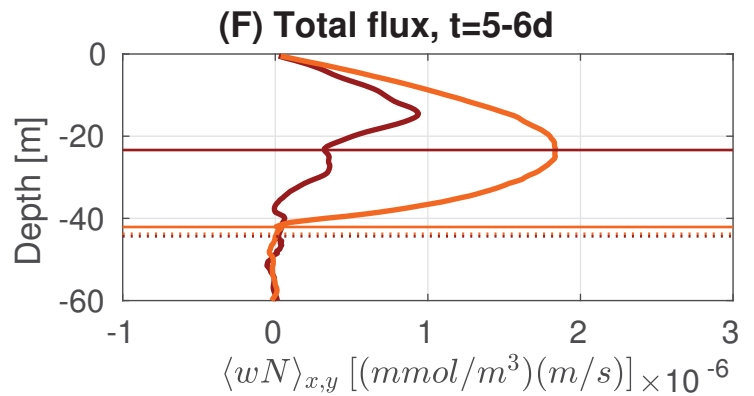
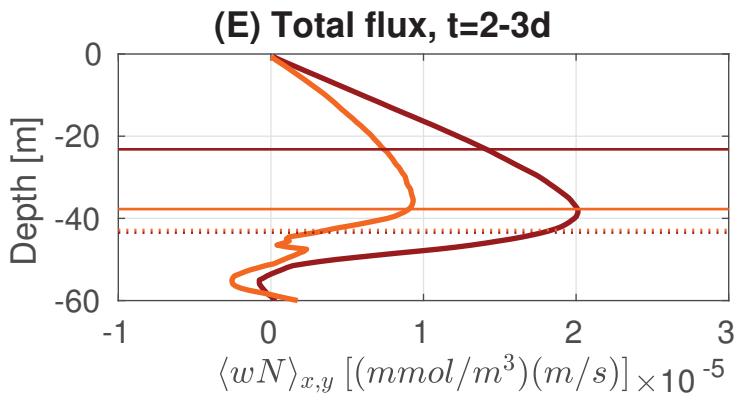
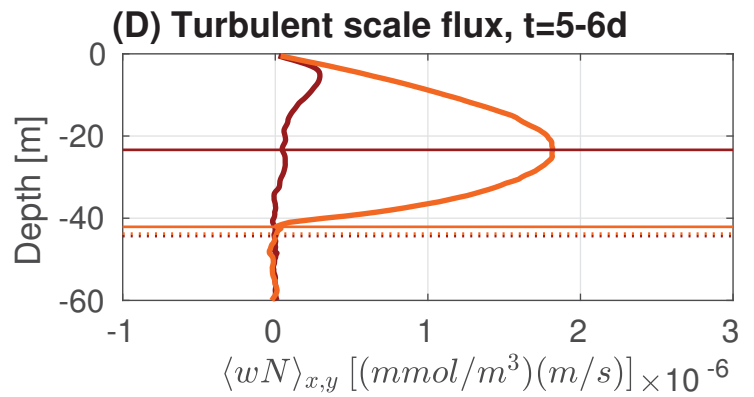
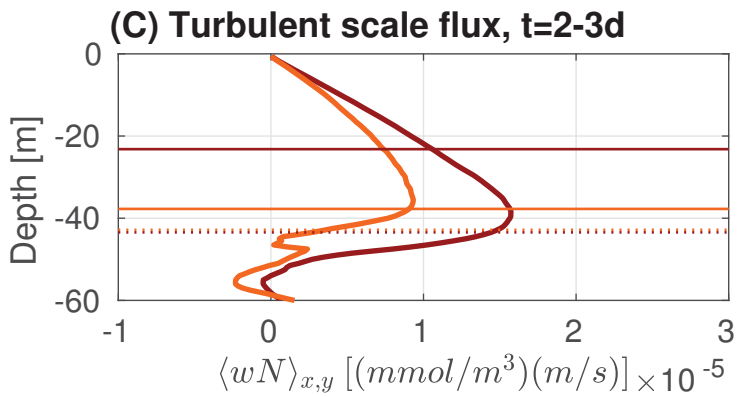
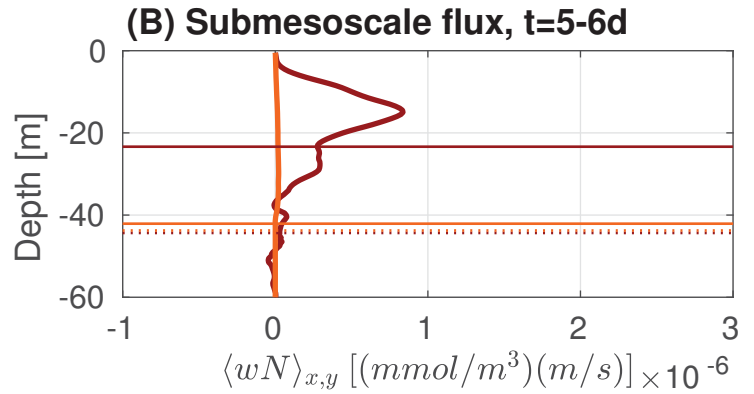
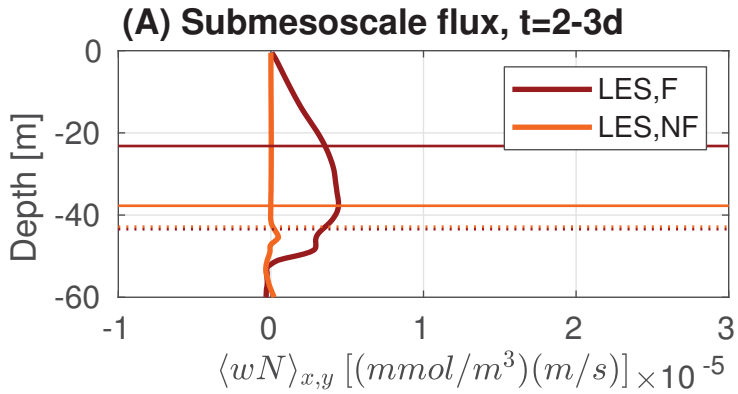
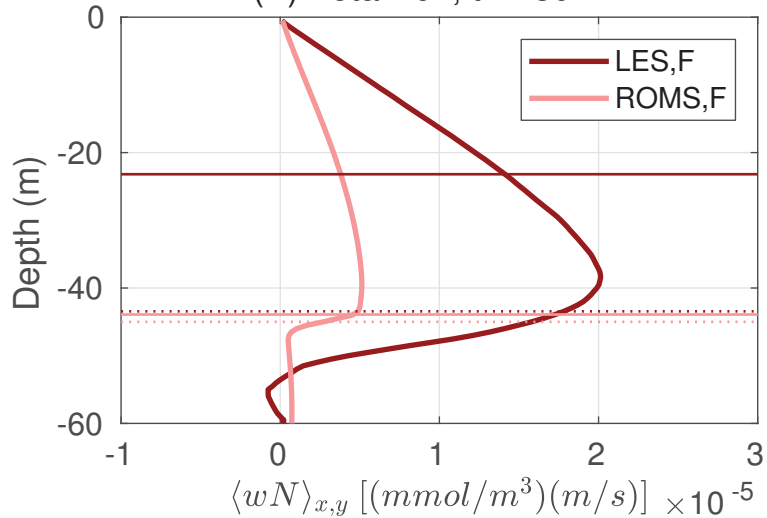
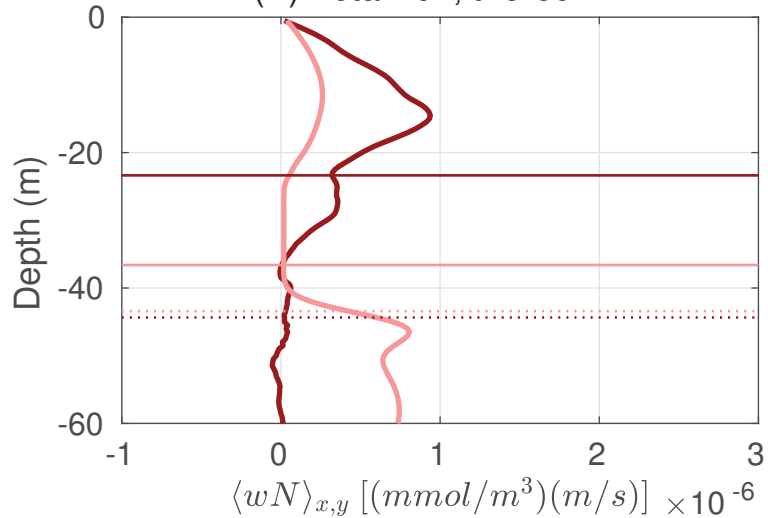


Figure 10.

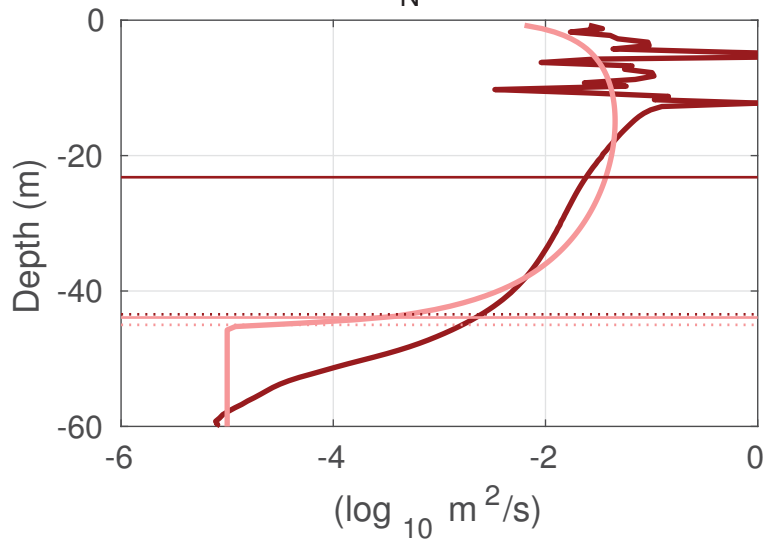
(A) Total flux, t=2-3d



(B) Total flux, t=5-6d



(C) κ_N , t=2-3d



(D) κ_N , t=5-6d

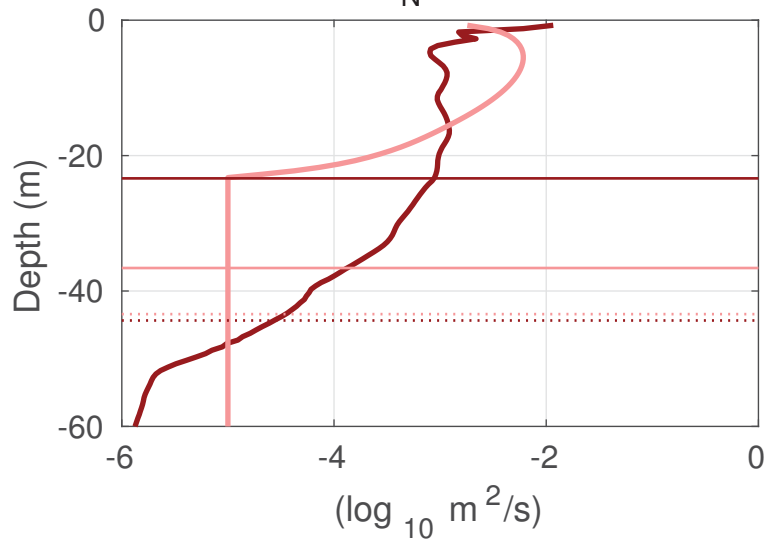
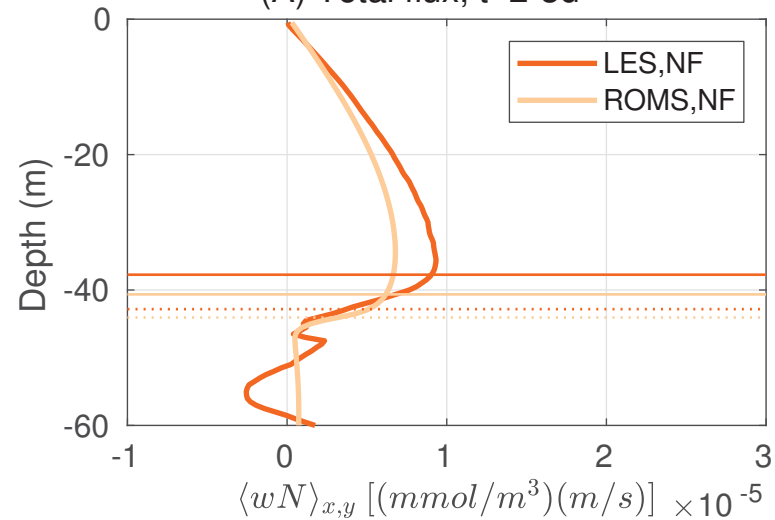
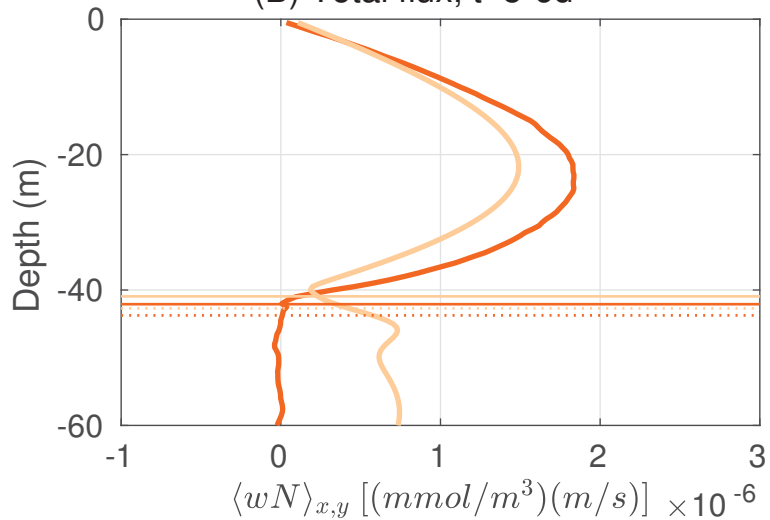


Figure 11.

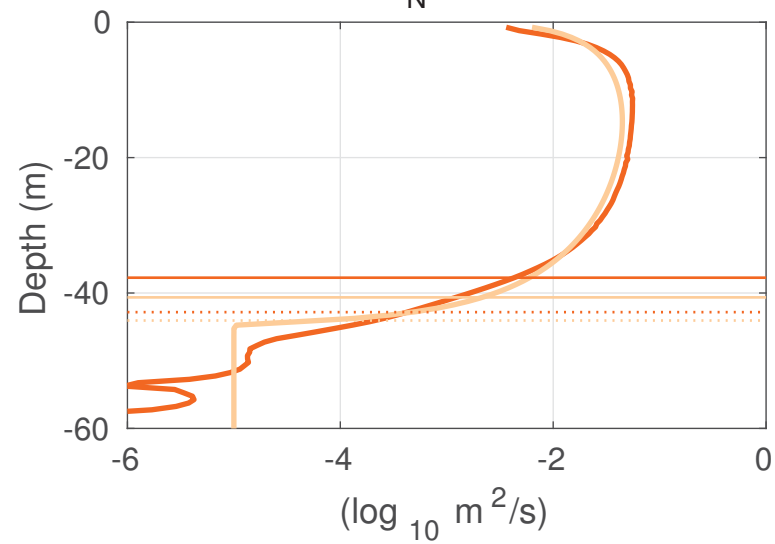
(A) Total flux, t=2-3d



(B) Total flux, t=5-6d



(C) κ_N , t=2-3d



(D) κ_N , t=5-6d

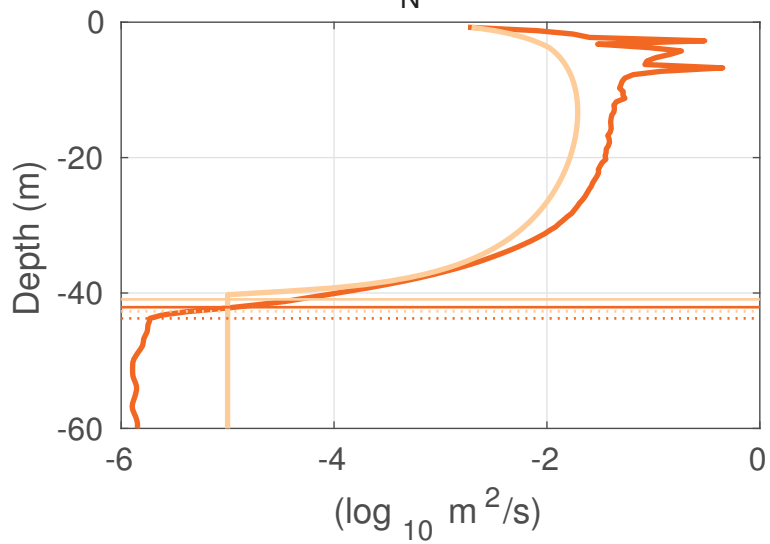


Figure 12.

