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1 **Competition between baroclinic instability and Ekman transport under**
2 **varying buoyancy forcings in upwelling systems: An idealized analog to the**
3 **Southern Ocean.**

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ABSTRACT

7 Coastal upwelling rates are classically determined by the intensity of the upper-ocean offshore
8 Ekman transport. But (sub-)mesoscale turbulence modulates offshore transport, hence the net
9 upwelling rate. Eddy effects generally oppose the Ekman circulation, resulting in so-called "eddy
10 cancellation", a process well studied in the Southern Ocean. Here we investigate how air-sea
11 heat/buoyancy fluxes modulate eddy cancellation in an idealized upwelling model. We run CROCO
12 simulations with constant winds but varying heat fluxes with and without submesoscale-rich
13 turbulence. Eddy cancellation is consistently evaluated with three different methods that all
14 account for the quasi-isopycnal nature of ocean circulation away from the surface. For zero heat
15 fluxes the release of available potential energy by baroclinic instabilities is strongest and leads,
16 near the coast, to nearly full cancellation of the Ekman cross-shore circulation by eddy effects,
17 *i.e.*, zero net mean upwelling flow. With increasing heat fluxes eddy cancellation is reduced and
18 the transverse flow progressively approaches the classical Ekman circulation. Sensitivity of the
19 eddy circulation to synoptic changes in air-sea heat fluxes is felt down to 125 m depth despite
20 short experiments of tens of days. Mesoscale dynamics dominate the cancellation effect in our
21 simulations which might also hold for the real ocean as the relevant processes act below the surface
22 boundary layer. Although the idealized setting overemphasizes the role of eddies and thus studies
23 with more realistic settings should follow, our findings have important implications for the overall
24 understanding of upwelling system dynamics.

25 **1. Introduction**

26 Along eastern boundary upwelling systems (EBUS) equatorward winds are responsible for a
27 thermally indirect Ekman circulation that brings heavy, deep, and generally nutrient rich water
28 towards the surface. Nutrient enrichment and new primary production have traditionally been
29 thought to linearly depend on the intensity of the alongshore wind stress component (Bakun
30 et al. 2010). This is what hundreds of studies concerned with EBUS ecosystem functioning have
31 implicitly assumed by relying on the so-called Bakun index (derived from alongshore coastal
32 winds, sometimes complemented by windstress curl information (i.e. the CUTI index by Jacox
33 et al. (2018)), as a proxy for planktonic food availability. However, the rate at which water is being
34 transferred from the ocean interior into the mixed layer may not be simply related to the Ekman
35 transport.

36 Quite recently, the contribution of alongshore pressure gradients have increasingly been ac-
37 counted for. In many upwelling sectors they can yield onshore velocities that counteract, or more
38 infrequently reinforce, the Ekman flow (Colas et al. 2008; Marchesiello et al. 2010; Ndoye et al.
39 2017; Jacox et al. 2018). In this study, we shed light on another potentially important effect: quasi-
40 balanced meso- and submesoscale turbulence is responsible for rectified eddy transport which
41 also tends to counteract the Ekman upwelling cell. Several studies have addressed this topic (e.g.
42 Lathuilière et al. 2010; Gruber et al. 2011; Colas et al. 2013) but the knowledge of the eddies
43 role in EBUS remains fragmented. In an attempt to remedy this the present study borrows more
44 directly from and draw connections with the Southern Ocean (SO) literature, in which eddy effects
45 have been more extensively investigated (Marshall and Radko 2003; Hallberg and Gnanadesikan
46 2006; Morrison et al. 2011). In the SO, transport in the direction transverse to the Antarctic
47 Circumpolar Current (ACC) axis computed in isopycnal coordinates is a small residual between

48 the mean Ekman-driven circulation due to wind pumping/suction (the so-called Deacon cell) and
49 eddy-induced transfers of mass driven by baroclinic instability processes (Marshall and Radko
50 2006; Marshall and Speer 2012). The eddy-induced circulation attempts to flatten isopycnal sur-
51 faces, reduce available potential energy, and therefore largely opposes the thermally direct Deacon
52 cell. As a result, it is widely accepted that the sole knowledge of the wind forcing does not provide
53 useful insight into the mean cross-frontal circulation of the SO. There air-sea buoyancy fluxes are,
54 somewhat counter-intuitively, much more informative than the wind forcing with respect to the
55 meridional tracer transport (Marshall 1997; Gent 2016).

56 Indeed, they constrain rates of fluid transfer across isopycnals in the mixed layer, hence also
57 subduction and obduction rates at the base of the mixed layer in steady state (subduction - resp.
58 obduction - being related to a convergence - resp. divergence - of diapycnal flux of fluid, Fig. 1).
59 Neglecting mixing in the ocean interior the time-averaged residual meridional transport is thus
60 entirely determined by the knowledge of air-sea buoyancy fluxes (Walín 1982; Marshall 1997). In
61 particular, with zero air-sea buoyancy fluxes the residual circulation in the surface layer (and in the
62 interior) must vanish, which means that eddies exactly cancel the Eulerian wind-driven circulation
63 (Fig. 1c).

64 The transverse across-shore circulation found in upwelling systems is often represented as a
65 2D cell as in Fig. 1a. Under the constraint of no divergence, fluid is being upwelled near the
66 coast to feed the offshore Ekman drift induced by alongshore wind stress. Naively considered as
67 a representation of a 2D, time averaged (or steady), and laminar (no eddy terms) upwelling, Fig.
68 1a poses a conundrum that we find a useful starting point. Consider the fluid transported by the
69 Ekman circulation in the mixed layer. The thermohaline structure being steady this fluid crosses
70 isopycnals as it moves offshore and a net buoyancy gain is needed that must exactly satisfy the
71 relation:

$$v_{ek} \cdot \partial_y \langle \bar{b} \rangle_{ml}(y) = \bar{B}_{ml}(y) \quad (1)$$

72 where v_{ek} is the wind-driven Ekman velocity vertically averaged over the mixed layer, $\langle \bar{b} \rangle_{ml}$ is the
 73 mean buoyancy in the mixed layer, where $b = -g\rho/\rho_0$ with g being the gravitational acceleration,
 74 ρ the potential density, and ρ_0 a reference density. \bar{B}_{ml} is the net buoyancy supply to the mixed
 75 layer, supposed to mainly result from air-sea exchanges sector where coastal upwelling is taking
 76 place (Fig. 2). The derivation of Eq. (1) is detailed in Appendix 1.

77 By analogy with the SO, we hypothesize and will demonstrate that: (i) in a 3D upwelling system
 78 eddy fluxes are responsible for a large contribution to the time and/or alongshore average buoyancy
 79 equation (ignored in Eq. (1), see Eq. (A3)) that cancels a fraction of the buoyancy advection by
 80 the Eulerian Ekman circulation (ii) the structure and intensity of the eddy and residual circulation
 81 is modulated by the buoyancy input $\bar{B}_{ml}(y)$. (i) is a key finding in a few past studies (Gruber et al.
 82 2011; Colas et al. 2012, 2013; Nagai et al. 2015), which is also true for tracers other than buoyancy
 83 (Lathuilière et al. 2010; Gruber et al. 2011) and will be subjected to careful considerations herein.
 84 (ii) is an original result of this study which is demonstrated by analysing the synoptic response of
 85 upwelling dynamics to buoyancy fluctuations.

86 One practical difficulty is that upwelling systems lack the periodicity attribute so the ACC
 87 framework and theory are not directly applicable. Net buoyancy input can arise from lateral fluxes
 88 at the northern and southern edges of the upwelling region and this is an important source of
 89 complication. In addition most upwelling systems have intense seasonal fluctuations, so time
 90 averaging over one or several years and widely distinct ocean states is needed to ensure that the
 91 neglect of the buoyancy tendency term in Eq. (1) is valid (see Appendix).

92 To make progress in the overall understanding of EBUS functioning, we thus consider an idealised
 93 upwelling configuration with periodic boundary conditions following Lathuilière et al. (2010). The

94 setting and methods are described in section 2. How these results can be interpreted in the context
95 of real ocean upwelling systems is extensively discussed in section 4. The paper finishes with our
96 conclusions in section 5.

97 **2. Model setup and methods**

98 *a. Idealized upwelling model configuration*

99 The ocean circulation model CROCO (Coastal and Regional Ocean COmmunity model,
100 www.croco-ocean.org) is used in an idealized upwelling configuration. CROCO is formulated
101 in terrain-following sigma coordinates. The size of the computational domain is 400 km (L_x) by
102 600 km (L_y) long in the alongshore (x) and cross-shore (y) direction respectively (Fig. 3). The
103 bottom topography $h(x,y)$ consists of a narrow shelf, a continental slope and a flat bottom over the
104 open ocean (Figs. 3). $h(x,y)$ is defined by the following analytic function:

$$h_i(x,y) = h_{max} \left[\frac{1}{2} \left(1 + \tanh \left(\frac{y - y_s}{L_s} \right) \right) \right] + \delta h_i(x) \quad (2)$$

105 with $h_{max} = 1960$ m, $y_s = 100$ km, $L_s = 50$ km. With these parameter values for Eq. (2) the
106 minimum depth at $y = 0$ km is 34 m and the shelf is about 50 km wide (Fig. 5c). Note that the
107 bottom topography is uniform in the alongshore direction except for a small perturbation $\delta h_i(x)$
108 with i being the index of the i^{th} ensemble run (see details in section 2b).

109 We use periodic boundary conditions in the alongshore direction. At the coastline the domain
110 is closed. For numerical simplicity a closed boundary is also present along the offshore side
111 of the domain. To limit the spurious effects of this western wall a sponge layer of width equal
112 to 96 km is included in which momentum and temperature are subjected to harmonic diffusion
113 and temperature is restored toward a background temperature profile $T_o(z)$ (see Eq. (3)). In all
114 simulations diffusion and viscosity coefficients increase smoothly via a cosine function from 0

115 m^2/s at $y = 506$ km to a maximum value of $600 m^2/s$ towards the closed boundary at $y = 600$
116 km (Fig. 3). For lateral momentum advection the default CROCO 3rd-order upstream biased
117 advection scheme with implicit diffusion is used. For temperature we use a 5th-order WENOZ
118 quasi-monotonic advection scheme to prevent temperature under/overshoots near sharp horizontal
119 fronts.¹ Spurious diapycnal mixing associated with this scheme cannot be entirely avoided but shall
120 have negligible effects in our study because 1) sigma-coordinates are quasi-horizontal over most
121 of the domain where the bottom is flat, 2) we are concerned with the upper 100-150 m of the water
122 column in which (physical) diapycnal mixing rates are moderately high 3) we use kilometre scale
123 horizontal resolution (except for some sensitivity runs; see Marchesiello et al. (2009) and Lemarié
124 et al. (2012) for relevant details on spurious diapycnal mixing). Free slip boundary conditions are
125 used at the onshore and offshore walls. Linear friction with a drag coefficient $d_{bottom} = 6 * 10^{-4}$
126 m/s is applied at the ocean bottom.

127 The temperature profile $T_o(z)$ is also used as an homogeneous initial condition for all simulations.
128 It is chosen to be typical of (sub)tropical EBUS. Specifically we choose the mean climatological
129 temperature profile of the CARS 2009 database (www.cmar.csiro.au/cars, Ridgway et al. (2002);
130 Condie and R. Dunn (2006)) for the month of January at $15^\circ S / 86^\circ W$ off Peru. For simplicity and
131 easier reproducibility we fit this profile with the following analytic function:

$$T_0(z_k) = T_a \left[1 + h_1 \log \left(\cosh \left(\frac{z_k - h_2}{h_1} \right) / \cosh \left(\frac{z_{k-1} - h_2}{h_1} \right) \right) / (z_k - z_{k-1}) \right] + T_b * \exp \left(\frac{z_k}{1000} \right) \quad (3)$$

132 with the following constants: $T_a = 4^\circ C$, $T_b = 15.5^\circ C$, $h_1 = 80 m$ being the vertical scale of the
133 thermocline width, $h_2 = -80 m$ being the depth of the thermocline center and k the index of
134 the vertical grid coordinate with $z(k = 1) = 1960$ and $z(k = 100) = 0 m$. The first part of Eq.

¹https://croco-ocean.gitlabpages.inria.fr/croco_oc/model/model.numerics.overview.html

135 (3) prescribes the intensified stratification within the upper 250 m as typically found in eastern
136 boundary systems. The second part can be interpreted as a weakly stratified main thermocline
137 present over the full water column (Fig. 3).

138 The Coriolis parameter ($f = -2\pi/T_f$) is set constant (no β effect) and corresponds to a f-plane
139 latitude of about 14.5°S resulting in an inertial period (T_f) of 2 days. Due to the absence of any β
140 effect, there is no westward propagation of Rossby waves. This is not of major importance as we
141 restrict the duration of our experiment to 120 days (including a spinup of 80 days) but is presumably
142 responsible for eddy kinetic energy (EKE) levels above typical values found in the real ocean for
143 EBUS sectors (Haney et al. 2001; Marchesiello et al. 2003; Gruber et al. 2011). The first baroclinic
144 Rossby Radius is around 57 km within our idealized setting. This choice of parameters makes our
145 results applicable to the Peruvian, West African and northern Benguelan upwelling systems but
146 the main findings should also be valid at higher latitude.

147 In order to resolve sharp vertical velocity and temperature gradients in the upper ocean, we
148 use 100 vertical sigma levels for all configurations. The vertical coordinate by Shchepetkin and
149 McWilliams (2009) are used with the vertical grid parameters $\theta_a = 6$, $\theta_b = 0$ and $hc =$
150 10. Upwelling systems typically have water depth between 4000 m and 6000 m offshore of the
151 continental slope. Here we use a reduced bottom depth of 1960 m in combination with the high
152 number of vertical levels results in a relatively high vertical resolution especially near the surface.
153 Over the upper shelf the vertical resolution ranges from 0.1 m near the surface to 1.2 m above the
154 bottom. Offshore the vertical resolution ranges from 0.7 m at the surface to 6 m at 100 m depth
155 and 112 m near the bottom.

156 In our reentrant upwelling channel no along-shore pressure gradient can a priori exist at the
157 system scale due to the periodic boundary conditions. Without an along-shore pressure gradient
158 no interior geostrophic onshore flow can be established. Feeding the surface Ekman flow would

159 therefore involve bottom Ekman transport, thereby limiting the realism of the model for most
 160 upwelling sectors (Lentz and Chapman 2004). Following Lathuilière et al. (2010) we alleviate this
 161 difficulty by adding a constant and horizontally homogeneous alongshore pressure gradient to the
 162 momentum equation driving an onshore return flow V_{PG} :

$$V_{PG}(z) = \frac{-\tau_x}{f(H_{PG} - H_{Ekman})} \quad \text{if } -H_{PG} < z < H_{Ekman}, \text{ and zero otherwise.} \quad (4)$$

163 with $H_{Ekman} = 45$ m and $H_{PG} = 200$ m. All simulations are spun up from the initial resting state
 164 $T(y, z) = T_0(z)$, where T_0 is defined in Eq. (3). The alongshore wind forcing (τ_x) increases slowly
 165 with a ramp from 0 to the maximum value of 0.075 Nm^{-2} over a time period of $t = 4$ days:

$$\tau_x(t) = 0.075 \left[\sin\left(\frac{\pi t}{8}\right) + 0.075 \cos\left(\frac{\pi t}{8}\right) \right] \quad (5)$$

166 After day 4 all runs are forced with constant wind stress forcing of 0.075 N/m^2 (Fig. 4a).

167 *b. Heat flux sensitivity experiments and ensemble runs*

168 An important aim of this study is to investigate the effects of varying atmospheric buoyancy
 169 (here heat) forcing on upwelling dynamics and more specifically buoyancy advection. A simple
 170 heat flux formulation is chosen based on restoring to a reference SST which equals $T_0(0)$ to be
 171 consistent with the initial conditions and the restoring at the offshore edge of the domain. The heat
 172 flux into the ocean is thus given by the following formula:

$$Q(x, y, t)_{heat} = \frac{1}{\rho_0 C_p} \cdot \frac{dQ}{dT} (SST_m(x, y, t) - T_0(0)) \quad (6)$$

173 with dQ/dT being the relaxation coefficient in W K^{-1} , and SST_m the model SST (Barnier et al.
 174 1995). ρ_0 corresponds to the reference density (1025 kgm^{-3}) and C_p is the heat capacity of
 175 seawater ($3985 \text{ Jkg}^{-1}\text{K}^{-1}$). During the spin up period until day 80 the relaxation coefficient dQ/dT
 176 is kept constant at $-25 \text{ Wm}^{-2}\text{K}^{-1}$ (Fig. 4b). To modulate heat fluxes we artificially modify dQ/dT .

177 After day 80 dQ/dT is adjusted so as to conduct three sensitivity runs: (i) no heat flux ($dQ/dT =$
178 0, hereafter NF), (ii) moderate heat fluxes (dQ/dT remains equal to $-25 \text{ Wm}^{-2}\text{K}^{-1}$, MF) and (iii)
179 enhanced heat fluxes ($dQ/dT = -50 \text{ Wm}^{-2}\text{K}^{-1}$, HF). MF experiments receive on average a heat
180 input of about 200 Wm^{-2} near the coast and 50 Wm^{-2} 200 km offshore (Fig. 5a). These values
181 are typical of net heat flux forcings in EBUS although large variability exists (Fig. 2).

182 Heat flux sensitivity experiments are carried out at meso- ($\Delta x = 8 \text{ km}$) and submesoscale
183 permitting ($\Delta x = 800 \text{ m}$) resolution. Additional information about the simulation settings is given
184 in table 1. Due to the turbulent nature of the ocean the eddy fluxes are not solely determined
185 by the forcings and are subjected to intrinsic stochastic variability (in addition to not being in
186 statistical equilibrium). The duration of each experiment is only 40 days, which is insufficient to
187 get statistically reliable averages. Thus we perform 12 ensemble runs for each experiment listed in
188 table 1. Ensemble runs differ from each other by small-amplitude perturbations $\delta h_i(x)$ added to the
189 alongshore invariant bathymetry, which are responsible for triggering instabilities of the upwelling
190 flow. Specifically, we use:

$$\delta h_i(x) = -\frac{1}{2} \sin(2\pi m_i \frac{x}{L_x}) - \frac{1}{2} \sin(2\pi n_i \frac{x}{L_x}) \quad (7)$$

191 with 12 factor pairs (m_i, n_i) being (4,7), (2,9), (6,5), (3,2), (5,2), (2,1), (1,4), (1,6), (1,8), (4,9),
192 (4,5), (4,3).

193 *c. Quantification of eddy effects and overturning streamfunction calculation*

194 Our periodic upwelling channel is oriented in the alongshore (x) direction. Within the whole
195 study we define the perturbation of a variable X as the deviation from the alongshore average (\overline{X}^x):

$$X'(x, y, z, t) = X(x, y, z, t) - \overline{X}^x(y, z, t). \quad (8)$$

196 Temporal averaging (denoted \overline{X}^t) is also frequently performed over the time window from day
 197 81 to day 120 that excludes spin up. Averaging over all realizations of a given ensemble run is
 198 denoted \overline{X}^e . $\overline{X}^{x,t,e}$ is abbreviated as \overline{X} .

199 Baroclinic instability (hereafter BCI) processes are known to be essential in upwelling systems
 200 as a source of meso- and submesoscale turbulence (Marchesiello et al. 2003). The strength of BCI
 201 will be classically evaluated by computing available potential energy (APE) release as the averaged
 202 covariance of the vertical velocity (w) and buoyancy (b) perturbations: $\overline{w'b'}$. Buoyancy is defined
 203 here as a function of temperature with $b = -g\alpha T/\rho_0$ with $\alpha = 0.24 \text{ kg}^\circ\text{C}/\text{m}^3$, $g = 9.81 \text{ m}/\text{s}^2$.

204 Our focus is on the description of the eddy role on the transport of buoyancy. In practice,
 205 we will use three different approaches to determine variants of residual overturning (transverse)
 206 streamfunctions: 1) the transformed Eulerian mean formulation of Held and Schneider (1999);
 207 isopycnal averaging of water volume flux at 2) constant across-shore distance (equivalent to aver-
 208 aging at constant latitude in the SO, Döös and Webb 1994); or at 3) constant depth (Nurser and
 209 Lee 2004a,b). Each of these three approaches has its own advantages and limitations as detailed
 210 below. Thus we apply all three of them to ensure that our main findings are robust, as previously
 211 done for the SO or in the atmosphere.

212 The first formulation explicitly involves the rate of APE release and is able to accommodate
 213 outcropping situations. Following Held and Schneider (1999) (see also Colas et al. 2013) we define
 214 an eddy streamfunction for the transformed Eulerian mean circulation as $\Psi_{Eddy}^{TEM}(y, z) = \overline{w'b'}/\overline{b_y}$
 215 where $\overline{b_y}$ is the mean cross-shore buoyancy gradient. Being computed at fixed location Ψ_{Eddy}^{TEM} can
 216 be easily compared to the Eulerian mean streamfunction

$$\Psi_{Mean}(y, z) = \overline{\int_z^0 v(x, y, z, t) dz}^{x,t,e} \quad (9)$$

217 Simple Eulerian averaging is well known to produce transport fields with unrealistically large
 218 diapycnal components (i.e. the flow associated with Ψ_{Mean} crosses the mean isopycnal sur-
 219 faces and is therefore inconsistent with the weakly diabatic nature of interior ocean circulation).
 220 $\Psi_{Res}^{TEM} = \Psi_{Mean} + \Psi_{Eddy}^{TEM}$ is one estimation of the residual circulation advecting mean buoyancy and
 221 accounting for eddy effects.

222 A more natural way to account for the quasi-adiabatic nature of oceanic motions is to perform
 223 averages using moving density surfaces (or here equivalently temperature surfaces) as vertical
 224 reference levels. To this end, we define the second formulation:

$$\Psi_{Res}^{iso-h}(y, T_0) = \overline{\int \int_{(x,z,t):T(x,y,z,t) \leq T_0} v(x, y, z, t) dx dz}^{t,e} \quad (10)$$

225 Heaving of the isopycnals (here also isotherms) by eddies is absorbed into Ψ_{Res}^{iso-h} which thus
 226 includes an eddy contribution. Ψ_{Res}^{iso-h} can be remapped back into depth space using the mean
 227 height of each isotherm as a function of cross-shore distance $\overline{z(T, y)}^{x,t,e}$ (see Nurser and Lee
 228 (2004a) section 2c). Doing so, an eddy contribution to Ψ_{Res}^{iso-h} can be defined as:

$$\Psi_{Eddy}^{iso-h}(y, z) = \Psi_{Res}^{iso-h}(y, z) - \Psi_{Mean}(y, z) \quad (11)$$

229 Near the surface, averaging and integration using temperature as a vertical coordinate poses
 230 some issues. Specifically, the remapping from temperature to depth space becomes problematic
 231 wherever the water column becomes well-mixed or nearly so. In addition, rare occurrences of warm
 232 water conditions at the surface strongly imprint on the form of Ψ_{Res}^{iso-h} and its remapping (Nurser
 233 and Lee 2004a). To remedy this, an alternative quasi-lagrangian approach, the third formulation
 234 here, involves isopycnal averaging at constant height but variable cross-shore distance (or latitude
 235 in the ACC context):

$$\Psi_{Res}^{iso-v}(T_0, z) = \overline{\int \int_{(x,y,t):T(x,y,z,t) \leq T_0} w(x, y, z, t) dx dy}^{t,e} \quad (12)$$

236 Ψ_{Res}^{iso-v} corresponds to the vertical transport of fluid colder than a temperature T at any given depth
 237 z . The proximity to the ocean surface has no effect on this method. On the other hand, isothermal
 238 averaging at constant z is strongly impacted near side boundaries where the structure of Ψ_{Res}^{iso-v} can
 239 be difficult to interpret (Nurser and Lee 2004a; Lee and Nurser 2012). See section 2c in Nurser
 240 and Lee (2004a) for more details.

241 The ocean sector where upwelling takes place is in close proximity to the eastern boundary and
 242 not situated very deep below the surface. Indeed we are typically interested in the depth range from
 243 below the mixed layer down to 150-200 m depth or less, i.e., the source region of the upwelled
 244 water. The interpretation of the transport streamfunctions Ψ_{Res}^{iso-h} and Ψ_{Res}^{iso-v} is thus subject to
 245 caution. Below we present and compare all approaches. The degree of resemblance between
 246 Ψ_{Eddy}^{TEM} , Ψ_{Eddy}^{iso-h} and Ψ_{Eddy}^{iso-v} will be considered as an indication of robustness.

247 3. Results

248 *a. General hydrography and circulation in the idealized upwelling system*

249 The general thermal structures and circulation features of the idealised upwelling configuration
 250 resemble those typical of real upwelling systems (Figs. 5, 6). The wind forcing results in an
 251 offshore Ekman transport in the upper 30 - 40 m of the water column (Fig. 5d). The Ekman cell
 252 (Ψ_{Mean}) is closed by a return flow reaching down to 200 m depth and has a maximum strength
 253 of $2.3 \text{ m}^2 \text{ s}^{-1}$ around 50 m depth (Fig. 5d). The Eulerian circulation is broadly consistent with
 254 isopycnal doming in the upper 150 m of the water column especially within 80 km from shore (Figs.
 255 5c,d). Mean surface temperatures as low as $14 \text{ }^\circ\text{C}$ are simulated at the shelf break (Figs. 5c,d).
 256 A frontal zone separates the upwelling waters from the open ocean where surface temperatures
 257 increase to $20 \text{ }^\circ\text{C}$ at 200 km offshore (Figs. 5c, d). An alongshore surface jet of about 0.4 m s^{-1}

258 in same direction as wind forcing develops over the shelf and upper slope (Fig. 5c). A subsurface
259 undercurrent in opposite direction hugs the continental slope with maximal velocities of 0.15 m s^{-1}
260 in 150 – 200 m depth (Fig. 5c).

261 The idealized upwelling solution also produces turbulent structures as found in upwelling systems:
262 mesoscale eddies, filaments and sharp submesoscale temperature fronts particularly at the finest
263 resolution (Fig. 6a). Typically an anticyclonic mesoscale warm core eddy forms at some point near
264 the shelf break with diameter of roughly 150 km and a sea level anomaly of about 10 cm (Figs. 6a,
265 b). Several smaller scale cyclonic vortices are formed around the anticyclone with cold filaments
266 at the edges of the mesoscale structures (Figs. 6a, b). Within the filaments the SST can drop
267 down to about 18°C . Submesoscale frontal dynamics are obviously richer at the higher resolution.
268 Downwards velocities of up to 100 m/day are found at the cold side of these fronts (e.g. at $x =$
269 150 - 200 km and $y = 150$ km in Fig. 6c). These vertical velocities are a crucial part of the eddy
270 flow that counteracts the Ekman transport of buoyancy as we investigate in detail in the following
271 chapter through three different approaches.

272 *b. Eddy effects under varying heat flux forcing*

273 The effect of varying heat flux forcing on the cross-shore circulation is investigated in this
274 section. Based on the SO literature our hypothesis is that changing heat fluxes affects the eddy field
275 with implications on the eddy heat fluxes, hence also on the eddy-induced and residual transverse
276 circulations. To test this, we compute the 3 different variants of eddy and residual streamfunctions
277 defined in section 2c. This is done for the submesoscale permitting simulations which represent
278 the turbulent processes more accurately. A detailed comparison between the simulations with 8
279 km and 800 m horizontal resolution is carried out in the next section 3c.

280 We first quantify the APE release by BCI to start looking into the eddy effect sensitivity to heat
 281 fluxes. The mean cross-shore distribution of $\overline{w'b'}$ shows maximum values above $1.8 \cdot 10^{-7} m^2 s^{-3}$
 282 at 20 m depth at around 30 km offshore during the NF case at submesoscale resolution (Fig.
 283 7a). Further offshore 200 km from the coast the maximum values decrease to values from 0.6-
 284 $0.9 \cdot 10^{-7} m^2 s^{-3}$ in 20 m depth (Fig. 7a). With increasing air-sea flux $\overline{w'b'}$ decreases (Fig. 7c)
 285 from maximum values $1.44 \cdot 10^{-7} m^2 s^{-3}$ in the NF case to $1.06 \cdot 10^{-7} m^2 s^{-3}$ in the HF case, when
 286 averaged between 16 to 112 km offshore. This represents a decrease of about 25%. The shoaling
 287 of the maximum $\overline{w'b'}$ (19 m in NF and 15 m in HF, Fig. 7c) is associated with a thinning of the
 288 mixed layer (27 m in NF and 23 m in HF, Fig. 7c) with increasing air-sea flux. Although $\overline{w'b'}$
 289 exhibits a marked decline with depth below the mixed layer, it remains enhanced down to about
 290 100 m depth within about 100 km from the shore, *i.e.*, where the tilt of the isopycnals is most
 291 pronounced (Fig. 7a). The modulation of BCI strength by air-sea fluxes reaches down to similar
 292 depths (compare $\overline{w'b'}$ in Fig.7c; at 75 m depth $\overline{w'b'}$ is reduced by about 50% in the HF compared
 293 to NF). Below 125 m depth the domain-averaged $\overline{w'b'}$ curves for all air-sea flux cases show almost
 294 no difference and vanish at around 200 m depth (Fig. 7c), consistently with the relatively shallow
 295 extension of mesoscale turbulence in EBUS (Capet et al. 2008).

296 We now turn to the effects of turbulence on buoyancy advection and start with the most dramatic
 297 NF case. Ψ_{Eddy}^{TEM} is shown in Fig. 7b. Its sign is systematically negative and it almost mirrors the
 298 structure of the mean streamfunction (Ψ_{Mean}) shown in Fig. 8a. The eddy streamfunction reaches
 299 values down to $-2.8 m^2 s^{-1}$, which is stronger (in opposing direction) than the mean Eulerian
 300 streamfunction ($2.3 m^2 s^{-1}$, Fig. 5d). Note that this characteristic of overcancellation is absent or
 301 attenuated in Ψ_{Eddy}^{iso-h} and Ψ_{Eddy}^{iso-v} (Figs. 8e and 9e). The recirculation confined in the mixed layer
 302 revealed by Ψ_{Eddy}^{TEM} (and to a lesser extent Ψ_{Eddy}^{iso-v}) is the signature of submesoscale-driven mixed
 303 layer restratification (Fox-Kemper et al. 2008). As expected this signature is improperly captured

304 by isopycnal averaging at constant cross-shore distance (Held and Schneider 1999). It is the main
305 point of disagreement between the three variants. In the ocean subsurface (at 50 - 100 m depth)
306 where we seek to understand the eddy effect on buoyancy advection, all three formulations agree
307 on patterns and intensity, even close to shore where the effect of the ocean boundary is supposed
308 to impact differently on each of them (see section 2c).

309 Mean-Eddy cancellation is manifest in Figs. 8i and 9i. In the upper 50 m at 200 km and even
310 down to 100 m closer to the shelf break the intensity of the residual circulation is smaller than that
311 of the mean Eulerian circulation by a factor 4 or more. Most importantly, advective feeding of
312 the surface layer with upwelling water vanishes. At greater depth (125 m) residual streamfunction
313 values above $1 \text{ m}^2 \text{ s}^{-1}$ are found away from the coast (Fig. 8i) where the tilt of the isopycnals and
314 BCI strength (Fig. 7a) is reduced. Note that the residual circulation below the surface layer is not
315 particularly better aligned with the isopycnals than the mean Eulerian circulation, contrary to what
316 is typically found in the SO (Karsten and Marshall 2002). There are two reasons for this. First,
317 our simulations are not in statistical equilibrium so that transient adjustments might be responsible
318 for apparent "mean" diapycnal flow over the period of analysis. The fact that our simulations
319 for the different air-sea flux cases exhibit similar residual circulations indicates that this reason is
320 secondary. Second, and more importantly, our entire region of interest is in close proximity to the
321 mean ML base and it is subjected to intense intermittent mixing. To substantiate this statement, we
322 show the maximum mixed layer depths reached for every simulation in Fig. 7. More specifically,
323 mixed layer depths between 60 and 100 m are reached in 5% of the time. Although our simulations
324 do not resolve all relevant processes, a transition layer with relatively elevated mixing levels is
325 found below the mixed layer (Large et al. 1994; Johnston and Rudnick 2009). This provides the
326 required diabatic forcing to accommodate a slow but non-zero residual circulation consistent with
327 upwelling to about 80 - 120 m depth which bifurcates offshore above this depth range (Figs. 8i,

328 9i). This circulation pattern is very similar for Ψ_{Eddy}^{iso-v} and Ψ_{Eddy}^{iso-h} which provides confidence in
 329 its robustness, despite the proximity of the ocean surface and coastline. By analogy with recent
 330 descriptions of the SO overturning circulation (Garabato et al. 2007; Silvester et al. 2014) one may
 331 see this feature as a short-circuiting of the coastal upwelling cell, resulting from the combination
 332 of eddy cancellation and mixing.

333 How these results are affected by the air-sea buoyancy forcings is described in the remainder
 334 of this section. In contrast to Ψ_{Mean} , the sensitivity of the eddy-induced and residual circulation
 335 to air-sea heat forcing is noticeable well below the mixed layer base (Figs. 9h, l). For instance,
 336 the lower part of the eddy cell weakens by over 50% (from $-1.3 \text{ m}^2 \text{ s}^{-1}$ in NF to $-0.55 \text{ m}^2 \text{ s}^{-1}$ in
 337 HF) at 70 km offshore and 75 m depth. Although a large degree of eddy-mean cancellation is still
 338 present at HF its residual circulation has recovered a structure that more closely resembles the mean
 339 Ekman circulation, with some streamlines unambiguously connecting the offshore-subsurface to the
 340 nearshore-surface sectors. Again no major difference is found between the sensitivities exhibited
 341 by Ψ_{Res}^{iso-h} and Ψ_{Res}^{iso-v} below the mixed layer (compare Figs. 8h,l and 9h,l). In the mixed layer,
 342 Ψ_{Res}^{iso-v} displays a reduction of the submesoscale restratification tendency consistent with theory,
 343 which confirms the superiority of this formulation near the ocean surface (Held and Schneider
 344 1999).

345 Subtle differences between simulations with different heat fluxes are more readily apparent in
 346 Fig. 10 where streamfunction values are shown at a particular location chosen because it lies on the
 347 (Eulerian) mean upwelling pathway (see 'x' symbol in Figs. 8 and 9). Fig. 10 also gives a sense
 348 of the dispersion among realizations from the ensemble runs. The behaviors of Ψ_{Eddy}^{TEM} and Ψ_{Eddy}^{iso-v}
 349 as the heat flux forcing changes are very similar. The largest incoherence between formulations is
 350 found for NF with Ψ_{Eddy}^{iso-h} being $\sim 20\%$ larger than Ψ_{Eddy}^{TEM} and Ψ_{Eddy}^{iso-v} . At the chosen location, the
 351 eddies cancel between 72% (Ψ_{Eddy}^{iso-v}) and 87% (Ψ_{Eddy}^{iso-h}) of the Ekman transport for NF while the

352 cancellation reduces to between 46% (Ψ_{Eddy}^{iso-v}) and 60% (Ψ_{Eddy}^{iso-h}) for MF (Fig. 10b). In the HF case
 353 all 3 methods show cancellation levels of around 40 %. Beyond modest quantitative differences the
 354 main robust conclusion concerns the sensitivity of the eddy and residual circulation to air-sea heat
 355 fluxes: just as the Walin theory would predict in a steady state situation, mean buoyancy advection
 356 by the Ekman indirect overturning circulation is strongly counteracted by the eddies when the
 357 air-sea heat fluxes are such that no or limited diapycnal flow occurs in the surface layer. Pending
 358 discussion in section 4, note though that the elevated degree of cancellation we obtain is for an
 359 upwelling configuration in which EKE is well above typical EBUS values.

360 *c. Resolution sensitivity*

361 Submesoscale frontal processes are well known to be instrumental in the dynamics of the mixed
 362 layer. Our focus is on the dynamics in a layer of upper ocean fluid situated below the mixed layer,
 363 where the role of the submesoscale has not been clearly established and may vary depending on
 364 the regime under consideration (Capet et al. 2016). The influence of resolution on our findings is
 365 thus investigated, by comparing analyses for simulations at $\Delta x = 800$ m with analogues at $\Delta x = 8$
 366 km.

367 Starting with $\overline{w'b'}$ is instructive. Inspecting Figs. 7a,c,d confirms that submesoscale processes
 368 (poorly represented in mesoscale runs) matter in the mixed layer and immediately below it. Below
 369 50 m depth release of available potential energy in meso- and submesoscale runs are indistinguish-
 370 able. Notably, in the present idealized EBUS regime, the factor ten difference in resolution does
 371 not dramatically alter the magnitude of $\overline{w'b'}$ in the mixed layer: for NF maximum values reach
 372 1.44 (resp. 1.15) $m^2 s^{-3}$ at $\Delta x = 800$ m (resp. $\Delta x = 8$ km), which corresponds to a decrease of
 373 about 25%. Similar decreases are found between simulations with medium and high heat fluxes.²

²This result is seemingly at odds with previous studies including Colas et al. (2012; their figure 10) where a 15-fold increase in horizontal resolution strongly increases APE to EKE conversion near the surface. In reality, note that the absolute change in mixed-layer $\overline{w'b'}$ when going

374 The mean cross-shore buoyancy gradient \bar{b}_y being weakly sensitive to resolution similar reduc-
375 tions of about 25% are found for the intensity of Ψ_{Eddy}^{TEM} in the mixed layer (compare Figs. 7b
376 and e). Below the mixed layer eddy streamfunction differences between meso- and submesoscale
377 simulations are small and presumably not significant given the dispersion within the ensemble
378 runs. Also note that eddy and residual streamfunctions exhibit small spatial variations in their
379 cross-shore-depth structures at different resolution, so the single point comparison in Fig. 10
380 should not be overinterpreted.

381 In short the eddy cancellation mechanism and its sensitivity to air-sea buoyancy forcing high-
382 lighted in the previous section for submesoscale-permitting simulations results mainly from
383 mesoscale turbulence effects. Although this may seem surprising given the shallow nature of
384 the problem under consideration this is consistent with the limited vertical extension of subme-
385 soscale turbulence, which is typically confined into the mixed layer in most oceanic regimes (see
386 (Capet et al. 2016) for a counter-example where Charney instability is present).

387 4. Discussion

388 The overarching objective of this research is to contribute to the understanding of the factors
389 influencing upper ocean enrichment in nutrients, primary production, and the fate of organic
390 matter in EBUS, including considerations on their temporal variability and spatial heterogeneity
391 (*e.g.*, differences between upwelling systems or upwelling sectors within a given EBUS). The results
392 suggest that i) the eddy-induced circulation can counteract (or cancel in the SO terminology) mean
393 advection by the Ekman circulation to a large degree ii) the level of cancellation can be subjected
394 to spatio-temporal modulations due to air-sea buoyancy flux variability which can lead to changes

from mesoscale to submesoscale rich model resolution is quite similar in Colas et al. (2012) and our work ($2.8 \cdot 10^{-8}$ in our NF case in Fig. 7c versus $4.5 \cdot 10^{-8}$ for the winter season off Peru in Fig. 10 of Colas et al 2012). The reason why the comparison is misleading at first sight is the great difference in subsurface mesoscale $\overline{w'b'}$ between our idealized simulations and a real upwelling system like the Humboldt system (see section 4).

395 in the degree to which eddy fluxes counteract the mean advection by the Ekman flow that is central
396 in coastal upwelling dynamics. This being said, the cancellation mechanism has been investigated
397 in a simplified numerical configuration which has three important limitations in terms of realism.

398 First and foremost, the simulations we analyse have EKE levels that significantly exceed those
399 found in the real ocean, by a factor 2 to 10 depending on which EBUS sector is considered. Elevated
400 EKE is an inherent consequence of our periodic channel simplification with which the β -effect
401 is not compatible. The absence of beta precludes the westward radiation of energy, for instance
402 through Rossby wave propagation or beta-drift of vortices (Carton 2010). To make progress despite
403 this important caveat we assume that the intensity of the eddy overturning streamfunctions scales
404 linearly with EKE. Marshall et al. (2012) and Mak et al. (2017) provide some support to do so
405 although obviously not in the specific context where the aim is to account for a missing β -drift.
406 Fig. 11 is produced based on this assumption, on the EKE-degree of cancellation pairs obtained
407 for each of our simulations, and on the EBUS EKE values reported by Gruber et al. (2011). See
408 caption of Fig. 11 for more details. It provides estimates for the amount of cancellation in the
409 four EBUS, going from 3 to 6% in the northern Benguela, to 10 to 30% in the California Current
410 System. Upper range values correspond to situations with ≈ 0 net air-sea buoyancy fluxes which
411 are not very common in this latter system when upwelling conditions prevail. These results are
412 consistent with the well-accepted view that the role of eddies in EBUS is not of leading-order (as
413 it is in the SO). But they point to the possible importance of eddy-induced circulations during
414 periods when air-sea buoyancy forcings are weak and perhaps more so when they are temporarily
415 negative (see Fig. 2). Note that this situation was not investigated. Eddy fluxes may also have an
416 appreciable time-averaged effect on tracer advection in sectors of the California Current System
417 where standing meanders yield intensified meso- and submesoscale activity (Centurioni et al. 2008;
418 Colas et al. 2013). An important caveat regarding this rescaling approach concerns the possibly

419 subtle relationship between the cross-shore EKE distribution and the cancellation of the Ekman cell
 420 by eddies. Take for instance the southern Benguela where offshore Agulhas rings lead EKE levels in
 421 excess of $400 \text{ m}^2 \text{ s}^{-2}$ just offshore of the continental shelf (Capet et al. 2008). Mesoscale structures
 422 produced remotely by the Agulhas retroflection or by the baroclinically unstable offshore flowing
 423 California Current contribute to EKE in the vicinity of the coastal upwelling sector. However these
 424 structures may not be an important local source of APE to EKE conversion, which is more naturally
 425 related to cancellation (see the formulation of Ψ^{TEM} above). Most generally, the local relationships
 426 between surface EKE, upper ocean $\overline{w'b'}$, and ultimately Ψ^{TEM} , involve various processes, including
 427 horizontal advection of EKE to which beta drift is a leading-order contributor (Chelton et al. 2011).
 428 Thus, cancellation estimates derived from Fig. 11 ignore, for instance, the fact that beta drift is
 429 more effective at low latitude which fundamentally alters the $\overline{w'b'}$ – EKE relationship. Regional
 430 ocean models can be very useful to make further progress (see below). Our study suggests that such
 431 model need not resolve submesoscale processes: eddy cancellation of the Ekman flow depends
 432 on subsurface turbulence which, our study demonstrates, remains overwhelmingly linked to the
 433 mesoscale despite the strong frontality of the environment under investigation.

434 Second, our study was based on numerical simulations that were analysed only over specific time
 435 periods of 40 days. Our findings on the cancellation process itself do not specifically depend on
 436 this particular time scale but the modulation of the cancellation in response to air-sea buoyancy
 437 fluctuations may. To explore this potential issue, we have computed the temporal evolution of the
 438 APE release rate $\overline{w'b'}^{x,e,S}(t)$ over the area S where the cancellation process is most important and
 439 defined by: $y_1 < y < y_2$ and $z_1 < z < z_2$ with $y_1 = 16 \text{ km}$, $y_2 = 112 \text{ km}$, $z_1 = 0 \text{ m}$ and $z_2 = 80 \text{ m}$.
 440 Comparison between the $\overline{w'b'}^{x,e,S}(t)$ evolutions after day 80 for the three heat flux forcings reveals a
 441 fast adjustment process that takes place over about 3-5 days (not shown). This time scale may seem
 442 short for a BCI process that we have previously associated with the mesoscale. We are presently

443 unable to offer additional insight into this. Following this initial adjustment, APE release remains
444 approximately stable until day 120 (not shown). From this we infer that the modulation of the eddy
445 cancellation process in response to environmental changes should occur on synoptic and longer
446 times scales (including seasonal and interannual) but not on shorter time scales, e.g., as a response
447 to a diurnal cycle. In all EBUS air-sea buoyancy fluxes exhibit fluctuations of large magnitude on
448 synoptic and seasonal time scales as illustrated in Fig. 2 for nearshore sites. Strong ocean cooling
449 events (-200 to -500 W m^{-2}) can occasionally occur in 3 of the 4 EBUS. Additional experiments
450 would be required to investigate the role of meso- and submesoscale processes in these relatively
451 rare situations where mixed layer depth can reach 100 m, which complicates comparisons between
452 cases. Net heat fluxes values are more typically in the range $0 - 150 \text{ W m}^{-2}$ (Fig. 2b), *i.e.*, consistent
453 with the forcings used in this study whose results should therefore be relevant to the real ocean,
454 provided that the rescaling of cancellation based on EKE levels proposed above is correct.

455 Third, for simplicity, wind variability has been ignored throughout the study whereas it is
456 an important aspect of upwelling dynamics. In the real ocean wind and air-sea buoyancy flux
457 variability may be correlated and combine to produce results distinct from those found in this study.
458 Such correlations vary from place to place (Send et al. 1987; Beardsley et al. 1998; Flynn et al.
459 2017; Lübbecke et al. 2019). For instance low winds can yield large (resp. weak) air-sea net heating
460 into the surface ocean because cooling through latent heat release is reduced (resp. because in some
461 regions like central California upwelling relaxations are associated with increased nebulosity and
462 reduced incoming solar radiation). By analogy with the SO functioning and in agreement with the
463 general understanding of baroclinic processes the intensity of the eddy cancellation mechanisms
464 shall roughly scale with that of the Ekman flow, *i.e.*, we expect compensation (Marshall and Radko
465 2003) to occur whereby increasing upwelling winds steepens the isopycnals which increases the
466 counteracting effect of eddies. In Fig. 11 the intensity of the eddy-induced circulation is expressed

467 as a fraction of the Ekman cell transport, for the particular wind strength that we chose. How
468 that fraction actually varies depending on the wind conditions and wind spatial structure (Capet
469 et al. 2004; Small et al. 2015; Bonino et al. 2019) would need to be investigated. Furthermore,
470 the use of a pure flux vs. restoring condition for the surface buoyancy boundary condition, as well
471 as the the timescale in the restoring case (Zhai and Munday 2014) may have an influence on the
472 sensitivity of the residual overturning to wind stress changes (Abernathy et al. 2011). This needs
473 to be investigated in future EBUS studies.

474 Finally, note that alongshore pressure gradient variability is another factor that can modulate
475 upwelling intensity on synoptic, seasonal and longer time scales (Werner and Hickey 1983; Huyer
476 et al. 1987; Colas et al. 2008; Marchesiello et al. 2010; McCabe et al. 2015; Jacox et al. 2018),
477 frequently in the sense of an upwelling reduction. In the northern Benguela and southern Ca-
478 nary current sectors where the eddy cancellation is particularly weak this effect and its temporal
479 variability may thus be difficult to discern. In the northern Humboldt the eddy cancellation is
480 presumably stronger but still modest in magnitude ($\sim 10\text{-}20\%$, Fig. 11). Diagnostics of the eddy-
481 induced circulation in realistic simulations for this upwelling sector tend to confirm our estimates.
482 Maximum eddy-induced streamfunction values for summer reported in Colas et al. (2013) (see
483 their Fig. 8) reach about $0.2 \text{ m}^2 \text{ s}^{-1}$, *i.e.*, about 15% of the Ekman transport for that season. The
484 CCS is the system where eddy cancellation is expected to be strongest and possibly cancel a large
485 fraction of the Ekman circulation in some circumstances (up to 30-40%, Fig. 11). It is also
486 the system where the manifestation of eddy fluxes has received the most observational attention
487 (Shearman et al. 1999; Pallàs-Sanz et al. 2010b), including on biogeochemical tracer dynamics
488 (Bograd and Mantyla 2005; Huyer et al. 2005; Pallàs-Sanz et al. 2010a). Two estimations of the
489 eddy cancellation strength in the California current system can be drawn from Nagai et al. (2015)
490 and Colas et al. (2013). Note that the two studies use very similar numerical configurations.

491 Counting streamfunction contours in Figs. 5a, b of Nagai et al. (2015) gives an annual mean eddy
492 cancellation intensity $\sim 2/8=25\%$ of the Ekman flow at 50 m depth and 100 km from shore for
493 central California. Slightly weaker but comparable values of 10 to 15 % are obtained in Colas
494 et al. (2013) for summer, when upwelling winds and air-sea heat fluxes are most positive. Both
495 estimates are within the range of values inferred from the present study.

496 **5. Summary and conclusion**

497 An idealized numerical model is used to study the effect of eddies on the (along-shore) mean
498 transport of buoyancy in a coastal upwelling. The eddy contribution to buoyancy advection tends
499 to counteract the advection by the Ekman transport, so as to limit the slope of the isopycnals.
500 The efficiency of this eddy cancellation process varies with the strength of the air-sea buoyancy
501 flux forcing: eddies are most effective at impeding the transport of buoyancy by Ekman currents
502 in situations where air-sea buoyancy fluxes provide no or limited warming of upwelled surface
503 waters drifting offshore, which allows more intense and deeper-reaching frontal conditions to be
504 produced. Limitations imposed by our idealized framework do not allow us to work in steady
505 state and limit the duration of our experiments to periods of tens of days. However, drifts in
506 the thermohaline structure of our simulations remain small and our results can be interpreted
507 using a steady-state Walin type reasoning (Marshall 1997): given the mean frontal thermohaline
508 structure of an upwelling system vanishing (or negative) air-sea buoyancy fluxes would imply that
509 the surface Ekman flow produces diapycnal transport of mass unless it is counteracted by eddy
510 transport, resulting in partial or total cancellation of the mean buoyancy advection by the Ekman
511 flow. Two limit cases and an intermediate situation are represented in Fig. 1. Our simulations fall
512 in between the intermediate case and the total cancellation case. Concerned by the methodological
513 limitations inherent to eddy flux estimations and descriptions we used 3 different standard methods

514 and found good agreement between them in our idealized setting. Identifying and quantifying eddy
515 cancellation in realistic model simulations is far more difficult, mainly because the along-shore
516 periodicity of our numerical configuration offers a much simpler framework for analyses. It also
517 limits the time-averaging and/or ensemble run size requirements in a context where stochastic
518 variability is important and can blur the role of eddies and its sensitivity to forcings. However
519 the simplifications we take advantage of have important implications in terms of model realism.
520 As thoroughly discussed in the previous section, the real ocean behaviour is expected to differ,
521 with much smaller eddy effects in terms of Ekman flow cancellation than the ones we reported.
522 Despite this important caveat, we think that the idealized posing on which the present work is
523 based is useful to develop intuition on the role of eddies in upwelling systems. In the same spirit,
524 a follow-up study attempting to gain insight into biogeochemical tracer dynamics is in progress.
525 More realistic EBUS studies on eddy cancellation would be useful to further clarify the eddy role
526 on tracer transport and distribution in EBUS, as pursued in the context of subtropical gyres by
527 Doddridge et al. (2016) and Doddridge and Marshall (2018).

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536 *Data availability statement.* The overall size of the model output of all ensemble runs used for
 537 this publication is 2.7 TB. This large size makes it impossible for us to provide a constant online
 538 data access. Instead the simulations are stored safely by the first author and can be made available
 539 on request.

540 APPENDIX

541 In this appendix, we mainly repeat the derivation of Marshall and Radko (2003) in the context of
 542 an upwelling system. We start with the advection-diffusion equation for the evolution of buoyancy

543 b

$$\frac{\partial b}{\partial t} + \mathbf{u} \cdot \nabla b = D(b) \quad (\text{A1})$$

544 where \mathbf{u} is the 3D velocity field and D is a 3D diffusion operator

545 We then introduce a Reynold averaging operator (alongshore averaging) to separate rapid turbu-
 546 lent fluctuations from the slower part of the flow

$$X = \bar{X} + X' \quad (\text{A2})$$

547 The equation of evolution for the low-passed buoyancy writes:

$$\frac{\partial \bar{b}}{\partial t} + \bar{\mathbf{u}} \cdot \nabla \bar{b} = -\nabla \cdot \overline{\mathbf{u}'b'} + \overline{D(b)} \quad (\text{A3})$$

548 Turbulence provides an additional term that we have placed in the rhs of Eq. (A3) but part of
 549 this term can actually be rewritten as advection of mean buoyancy by an eddy induced velocity
 550 field. Following Held and Schneider (1999) or Marshall and Radko (2003), and taking advantage
 551 of the alongshore periodicity of our upwelling channel (which makes the eddy flux component
 552 in that direction $\overline{u'b'}$ irrelevant), the eddy flux is decomposed into an along-isopycnal component
 553 plus a leftover as follows (see Colas et al. 2013 for an alternative decomposition with purely

554 along-isopycnal and diapycnal components) :

$$555 \quad \left(\overline{v'b'}, \overline{w'b'} \right) = \left(\overline{w'b'}/s_{\bar{b}}, \overline{w'b'} \right) + \mathbf{F}_b \quad (\text{A4})$$

556 where $s_{\bar{b}} = -\frac{\bar{b}_y}{\bar{b}_z}$ is the isopycnal slope and $\mathbf{F}_b = \left(\overline{v'b'} - \overline{w'b'}/s_{\bar{b}}, 0 \right)$ is an associated diapycnal eddy
557 flux component. (A4) can be rewritten with the help of an eddy-induced vector streamfunction

$$\Psi_{Eddy}^{TEM} = \frac{\overline{w'b'}}{\bar{b}_y} \mathbf{i}, \quad (\text{A5})$$

558 where \mathbf{i} is the unit vector in the alongshore direction as:

$$559 \quad \left(\overline{v'b'}, \overline{w'b'} \right) = \Psi_{Eddy}^{TEM} \times \left(\bar{b}_y, \bar{b}_z \right) + \mathbf{F}_b. \quad (\text{A6})$$

560 Finally, taking the divergence of (A6), (A3) can thus be rewritten:

$$\frac{\partial \bar{b}}{\partial t} + (\bar{v} + v^*) \bar{b}_y + (\bar{w} + w^*) \bar{b}_z = \overline{D(b)} - \nabla \cdot \mathbf{F}_b \quad (\text{A7})$$

561 with the so-called "bolus" velocities defined as

$$562 \quad (v^*, w^*) = \nabla \times \Psi_{Eddy}^{TEM}, \quad (\text{A8})$$

563 In the situation of a 2D laminar and steady-state upwelling where the Ekman circulation is the
564 only flow component, Eq. (A7) can be written:

$$\bar{v} \bar{b}_y + \bar{w} \bar{b}_z = \overline{D(b)} \quad (\text{A9})$$

565 Integrating vertically over the mixed layer in which the buoyancy gradient is supposed to be
566 horizontal (no vertical stratification) and independent of depth, and neglecting heat, we find:

$$V_{ek} \left(\bar{b}_y \right)_{ml} = \overline{B_{ml}}(y) \quad (\text{A10})$$

567 where \overline{B}^{ml} is the net buoyancy input to the mixed layer and V_{ek} is the Ekman transport. This
568 relationship may seem to exert a strong constraint on the mean upper ocean thermohaline structure

569 of upwelling systems because given the wind and buoyancy forcings $\left(\overline{b}_y\right)_{ml}$ would need to adjust
570 so that the left- and right-hand side can match. Over relatively short study periods (40 days for the
571 analyses we carried out) temporal tendency could contribute to the balance but we do not find this
572 term to be important. In contrast we find that the eddies play an important role so that the neglect
573 of the bolus velocity in (A9) is invalid (this is also the case in Southern Ocean). In real upwelling
574 systems the eddy terms are not as strong (see Sec. 4) but the upper ocean buoyancy balance can
575 also involve mean alongshore advection term because of lack of periodicity.

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797 pii/S1463500314001358](https://www.sciencedirect.com/science/article/pii/S1463500314001358).

798 **LIST OF TABLES**

799 **Table 1.** Model run acronyms, horizontal resolution, number of grid cells, time steps,
800 duration, ensemble number, heat flux parameter dQ/dt and wind stress. 40

801 TABLE 1. Model run acronyms, horizontal resolution, number of grid cells, time steps, duration, ensemble
 802 number, heat flux parameter dQ/dt and wind stress.

Acronym	Δx , m	grid cells (x, y, z)	time step, s	duration, days	ensembles	dQ/dT , $W m^{-2} K^{-1}$	wind stress, $N m^{-2}$
Submesoscale No Flux, NF	794	504, 756, 100	80	80	12	0	0.075
Submesoscale Medium Flux, MF	794	504, 756, 100	80	80	12	- 25	0.075
Submesoscale High Flux, HF	794	504, 756, 100	80	80	12	- 50	0.075
Mesoscale No Flux, NF	8000	50, 75, 100	540	80	12	0	0.075
Mesoscale Medium Flux, MF	8000	50, 75, 100	540	80	12	- 25	0.075
Mesoscale High Flux, HF	8000	50, 75, 100	540	80	12	- 50	0.075

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804	Fig. 1.	Schematic showing the effect of atmospheric buoyancy forcing ($\overline{B_{ml}}$) on the cross-shore circulation in upwelling regimes. The offshore Ekman transport (v_{ek}) is shown in black arrows. The strength of eddy cancellation is sketched with colored arrows with blue representing colder upwelled water and red warmer offshore waters. Case (a) represents the traditional view of upwelling systems, (b) is consistent with the typical Southern Ocean conception including eddy cancellation and (c) represents an extreme situation where vanishing buoyancy flux and intense eddy activity lead to full cancellation. Note that the 3 idealized cases of this study fall in between b and c. The coast is located in $y = 0$
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812	Fig. 2.	Annual mean net buoyancy flux for 2009 to 2013 period from TropFlux (Praveen Kumar et al. 2012) (a). The four eastern boundary upwelling systems (EBUS): California, West Africa, Humboldt and Benguela are marked with black, red, blue and green squares (a). Annual mean (squares) heat fluxes, standard deviation (crosses) and maximum and minimum (dots) for the four EBUS Systems for the 2009 to 2013 period (b). Daily net buoyancy fluxes for each EBUS system for 2012 (c). The TropFlux data is produced under a collaboration between Laboratoire d'Océanographie: Expérimentation et Approches Numériques (LOCEAN) from Institut Pierre Simon Laplace (IPSL, Paris, France) and National Institute of Oceanography/CSIR (NIO, Goa, India), and supported by Institut de Recherche pour le Développement (IRD, France). TropFlux relies on data provided by the ECMWF Re-Analysis interim (ERA-I) and ISCCP projects. The vertical dashed black lines in b and c indicate the parameter space ($0 - 80 W/m^{-2}$) explored in this idealized study.
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824	Fig. 3.	Schematic of idealized eastern boundary upwelling model configuration. Model setup and schematic were inspired by Lathuilière et al. (2010). The model grid consists of a shallow shelf, a continental slope and a flat bottom. The Coriolis parameter (f) is constant and corresponds to a latitude of $14.5^{\circ}S$. The surface ocean is forced with a constant wind stress of $0.075 Nm^{-2}$ and varying heat fluxes (details in section 2 b). A sponge layer exists offshore with enhanced diffusivity and viscosity coefficients. There full depth temperatures are restored to the initial temperature profile $T_o(z)$ (see Eq. (3)). An alongshore pressure gradient is prescribed in the upper 200 m which drives an onshore flow balancing the offshore Ekman transport.
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833	Fig. 4.	Domain averaged wind stress (a) and heat flux (b) forcing during the model spin-up (day 0 to 80) and the three different heat flux experiments (day 80 to 120). The NF, MF and HF experiments are shown in black / gray, dark / light blue and dark / light red for the submesoscale ($\Delta x = 800m$) and mesoscale ($\Delta x = 8 km$) runs respectively. Single runs and ensemble mean are shown in thin and thick lines respectively.
834		46
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838	Fig. 5.	Temporal-mean alongshore-averaged heat flux (a), zeta (b), alongshore (c) and crossshore circulation (d) in reference simulation (MF) averaged over the time of the experiments (day 81 to day 120). Isothermes are contoured in white (c, d). Eulerian mean streamfunction is shown in black contours (d).
839		47
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842	Fig. 6.	Sea surface temperature (SST, in $^{\circ}C$) (a, b) and vertical velocities in m/day (c, d) in 20 m depth at day 110 for medium heat flux forcing for submesoscale ($\Delta x = 800 m$, a, c) and mesoscale ($\Delta x = 8 km$, b, d) horizontal resolution respectively. Sea level anomaly (zeta) in cm is shown in black contours.
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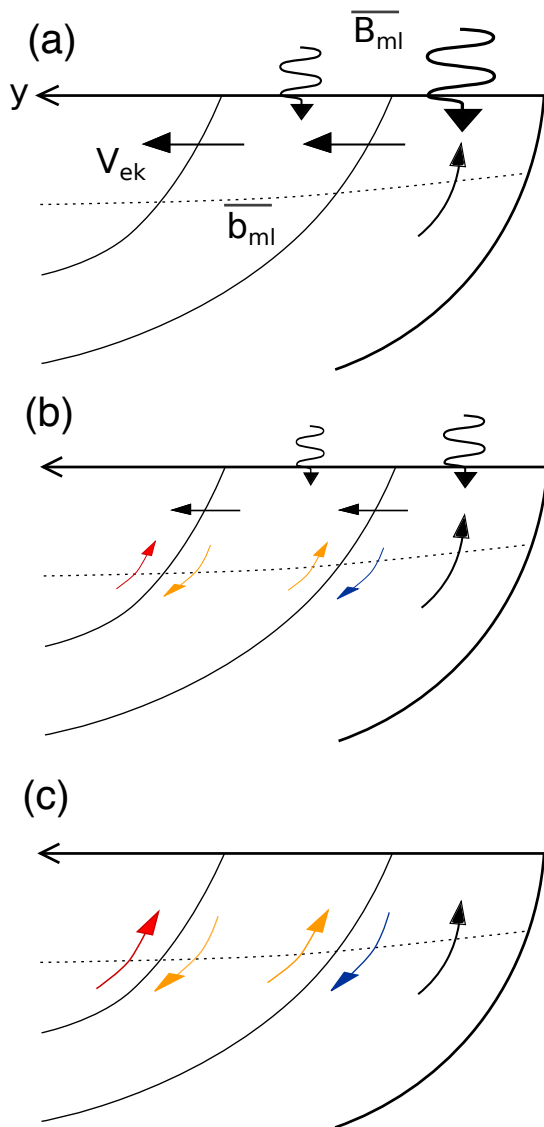
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 852 between 16 to 112 km with filled circles, squares and triangles in c respectively. 49

853 **Fig. 8.** Eulerian Ψ_{Mean}^{iso-h} (a, b, c, d), eddy Ψ_{Eddy}^{iso-h} (e, f, g, h) and residual Ψ_{Res}^{iso-h} (i, j, k, l) stream-
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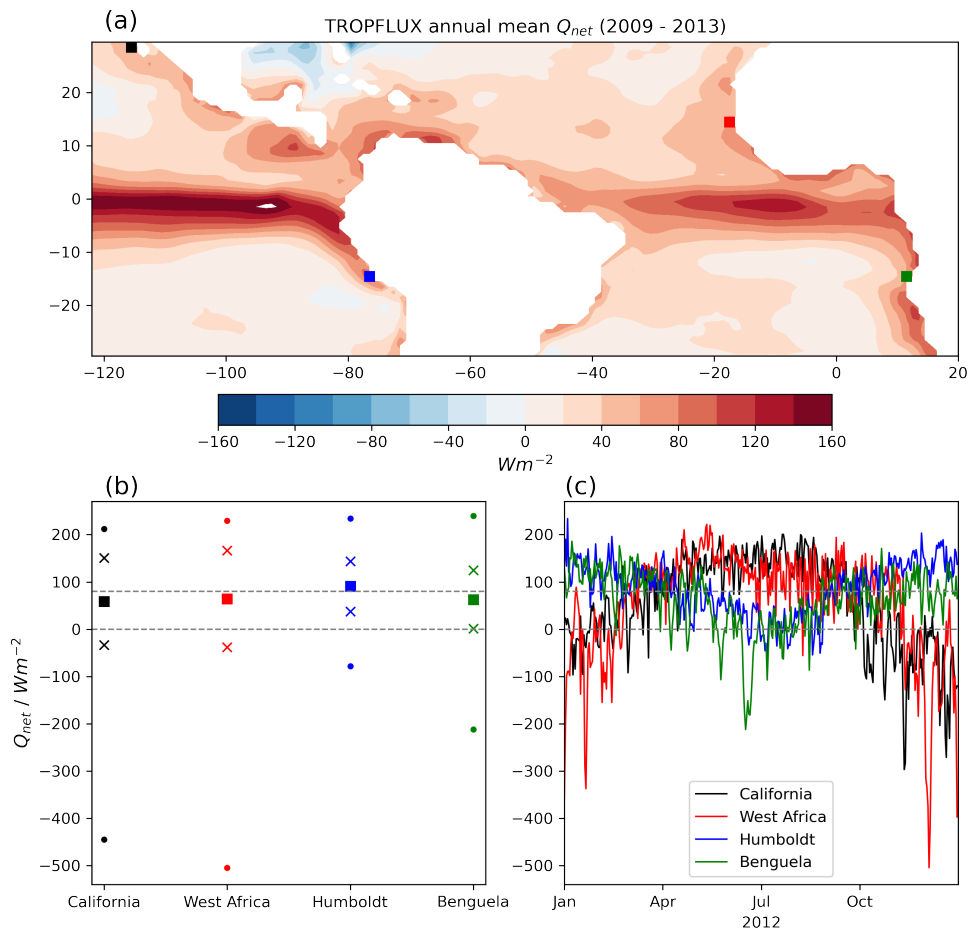
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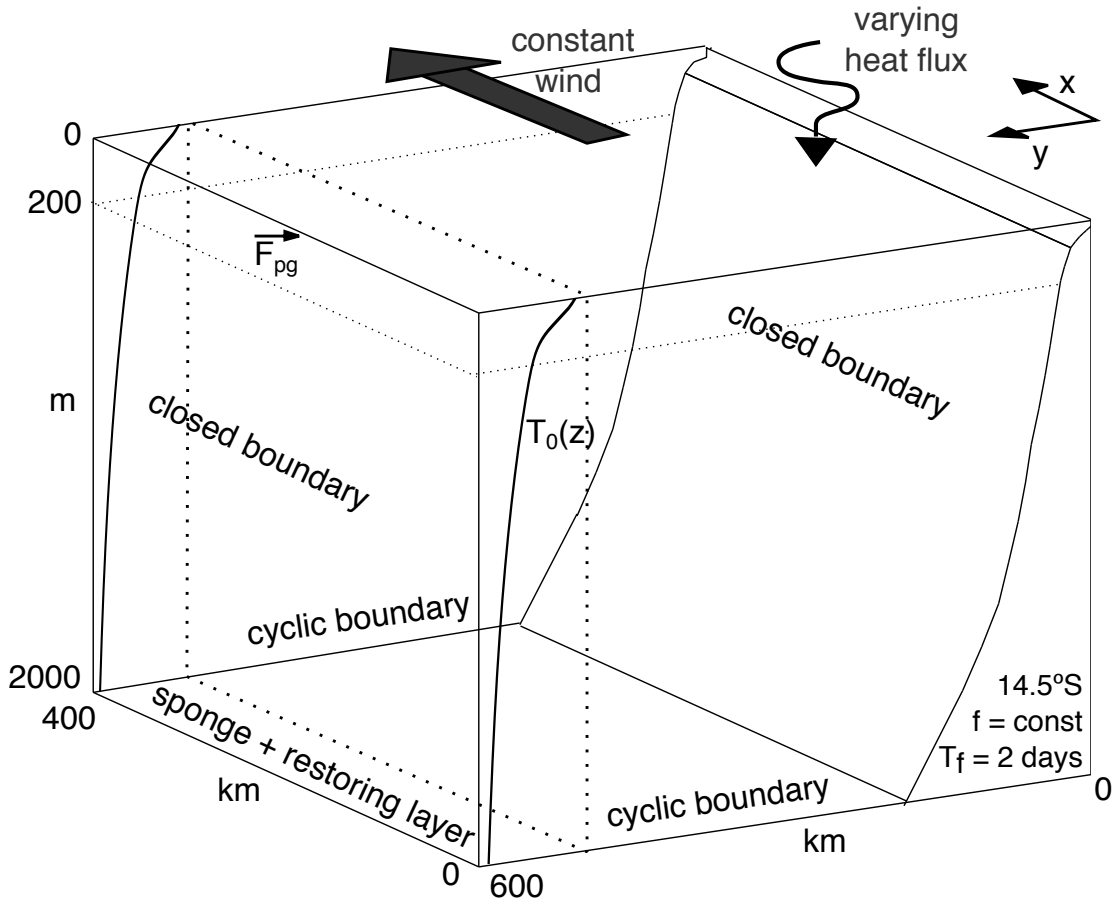
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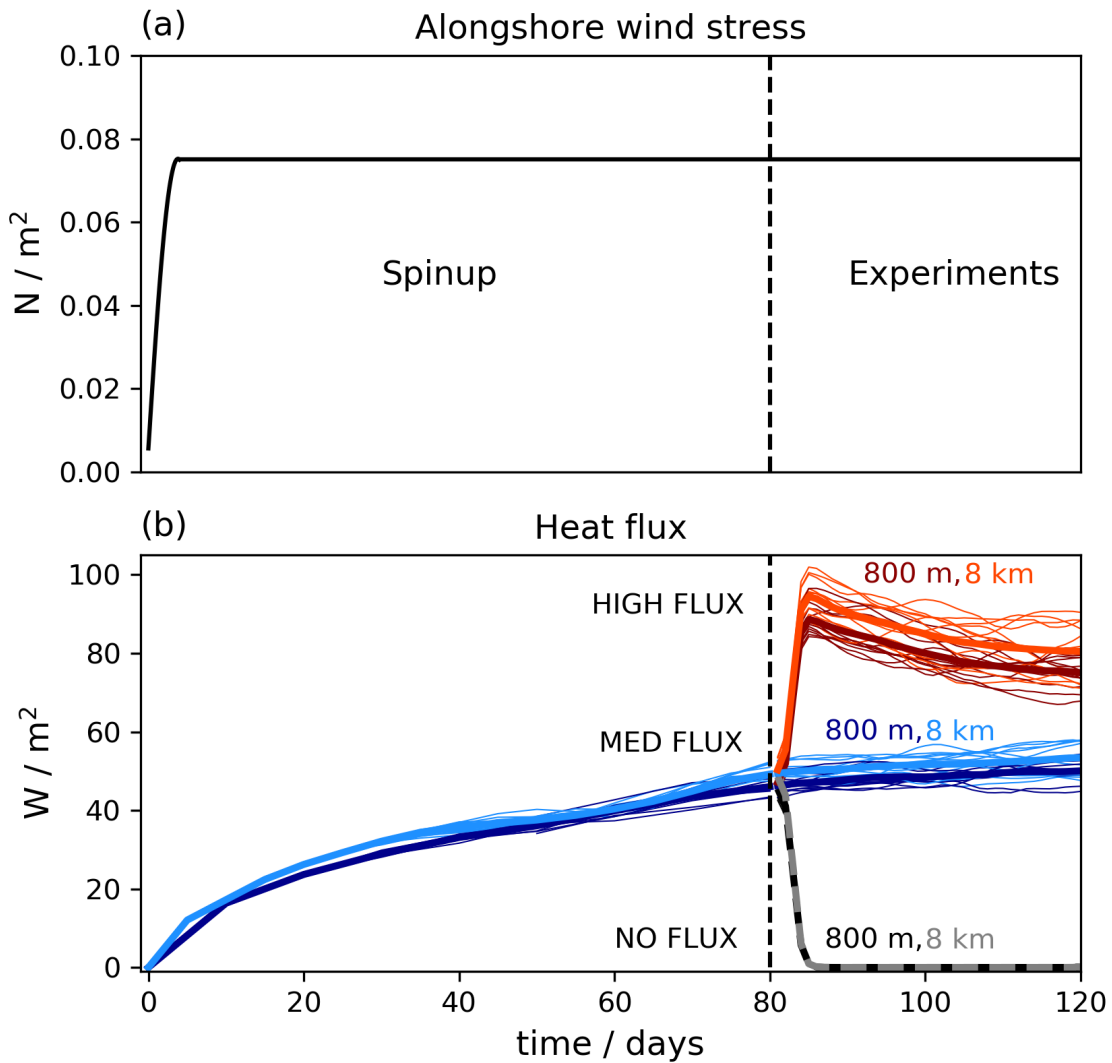
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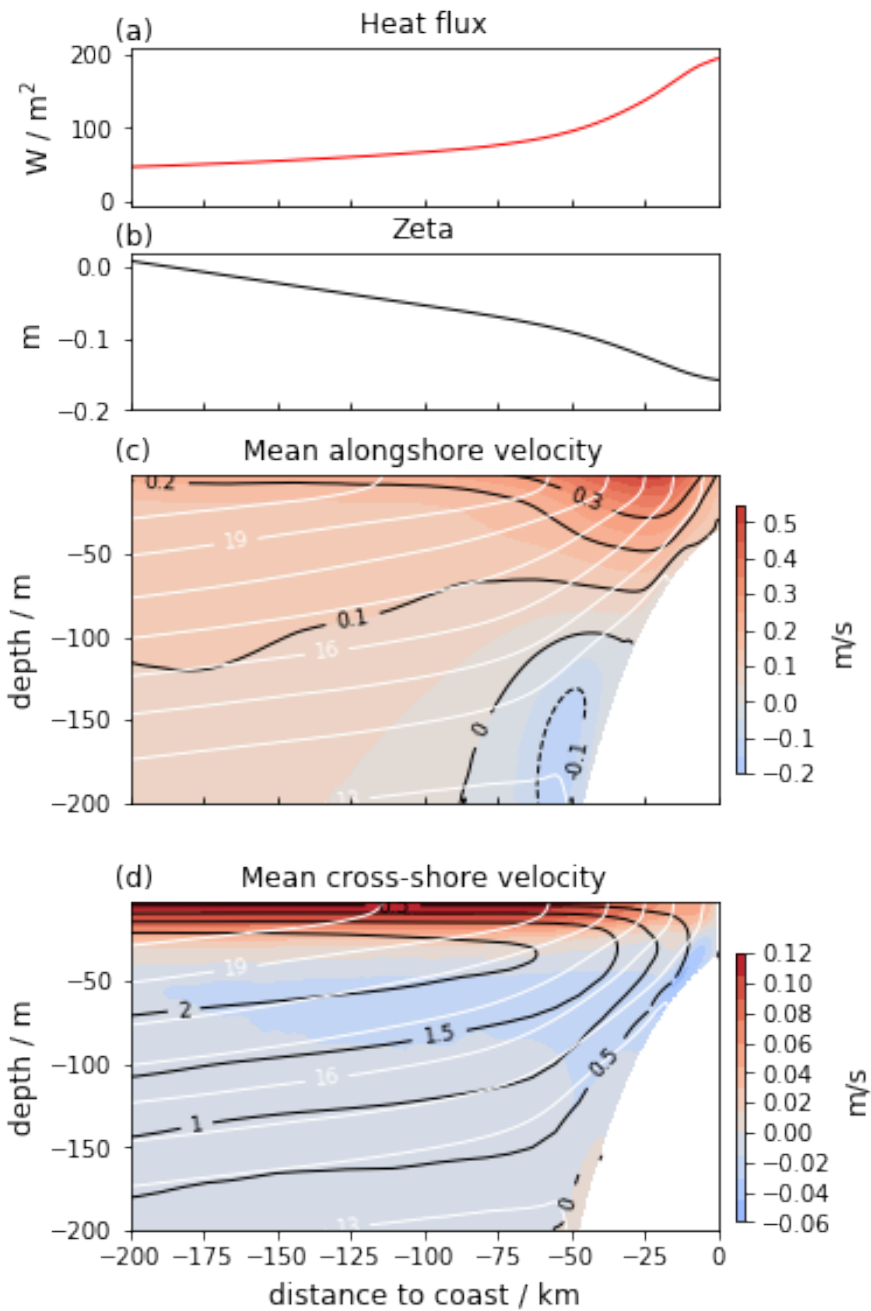
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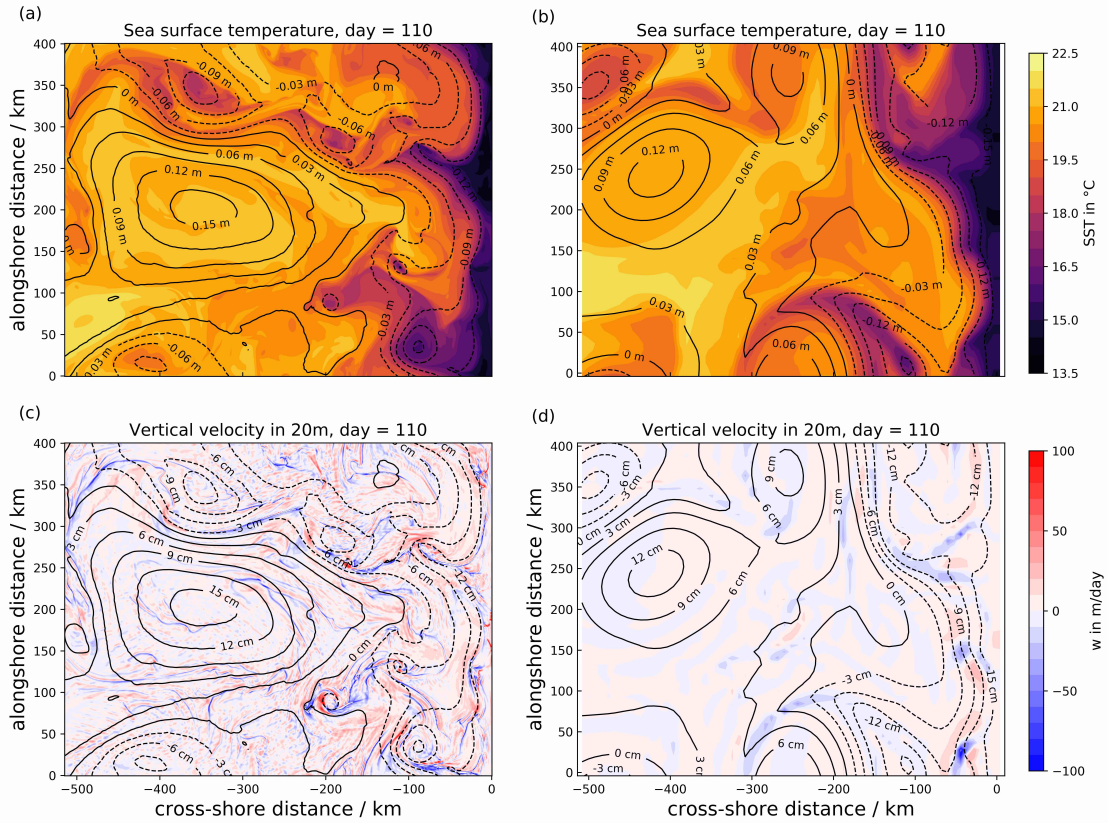
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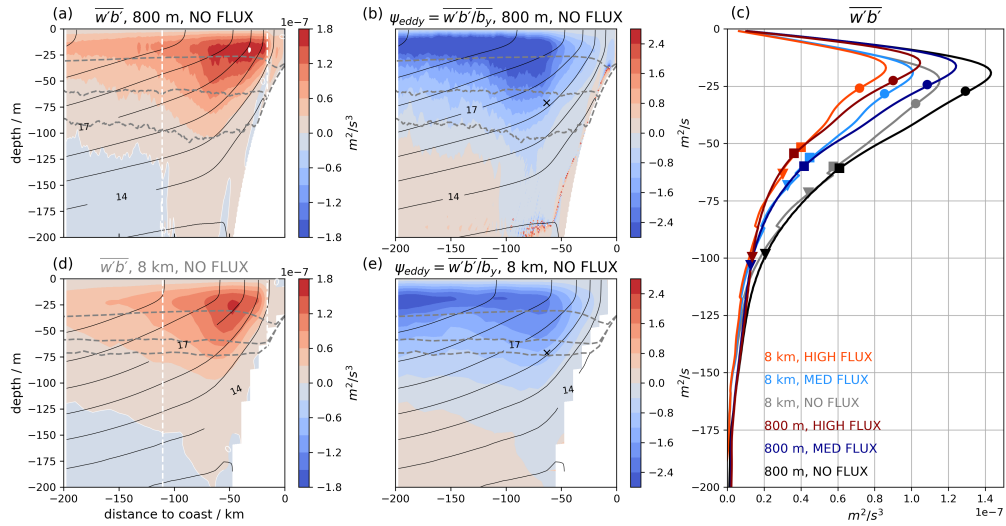
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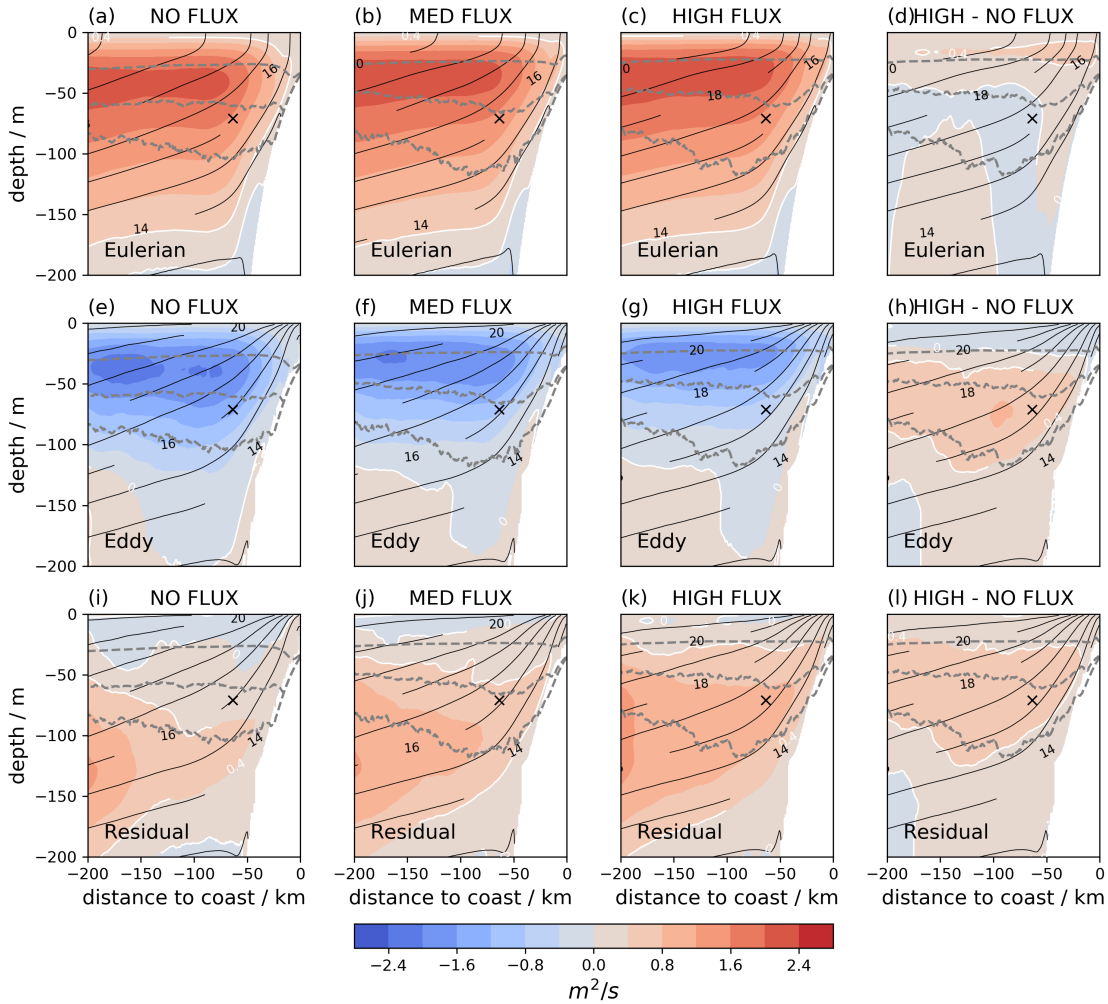
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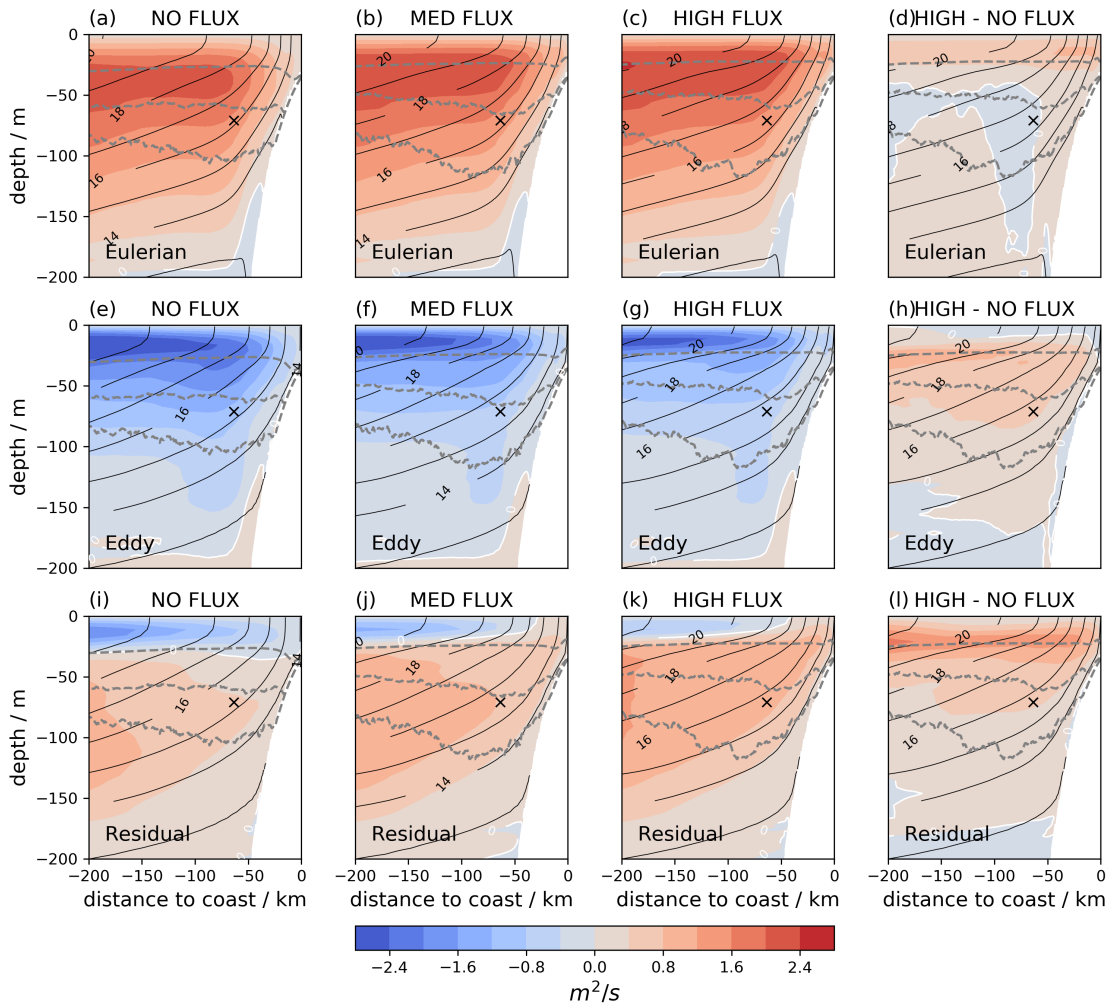
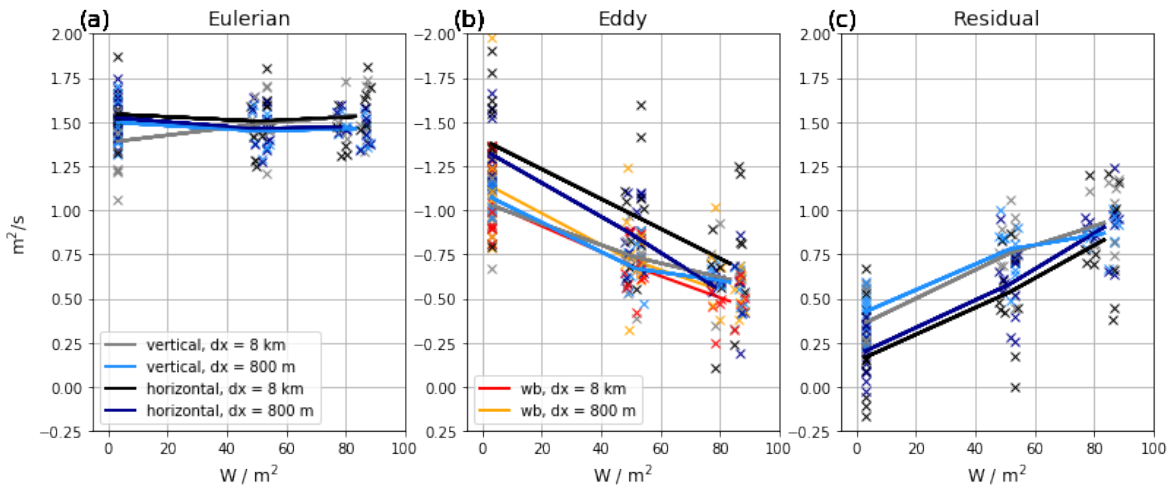
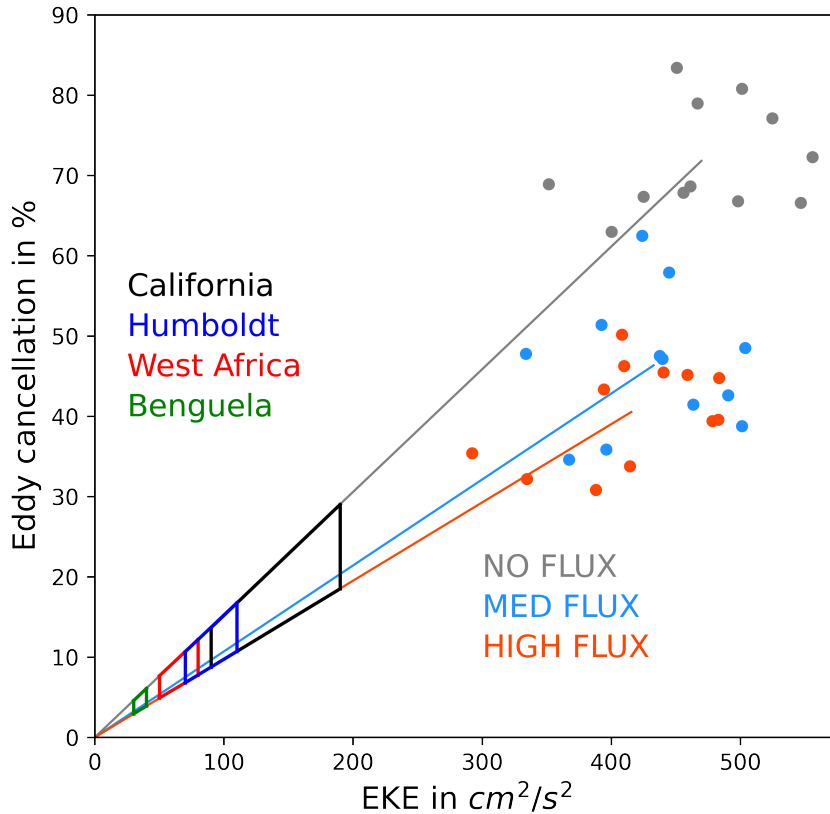


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