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The influence of the land-sea breeze on coastal upwelling systems: locally forced vs internal wave vertical mixing and implications for thermal fronts

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Abstract :

Land-sea breeze forcing near a land boundary drives both a locally forced response and an associated offshore propagating internal wave response, the effects of which can be difficult to separate. These processes enhance vertical mixing near the critical latitude for diurnal-inertial resonance (30° N/S), and are a feature of all four major eastern boundary upwelling systems. Here, we employ 1D- and 2D-vertical model configurations forced by a land-sea breeze to quantify the relative contributions of the locally forced and internal wave responses to surface currents and vertical mixing, and test sensitivity to latitude and bottom slope. We further include a sub-inertial alongshore wind to consider the role of the land-sea breeze in the context of upwelling systems. At the critical latitude, the internal waves generated via thermocline pumping near the land boundary are evanescent (in agreement with theory) and largely absent 50 km offshore. The internal waves are shown to contribute to vertical mixing, which can be 20% greater than that due to the forced response alone, further deepening the surface Ekman boundary layer. This deepening reduces the sub-inertial offshore advection of surface waters, thereby retaining the upwelling front closer to the land boundary and driving a net warming of the nearshore surface waters. Cross-shore horizontal oscillations of the upwelling front generated by the land-sea breeze drive strong diurnal variability in sea surface temperature, in agreement with observations from a cross-shore mooring array in the southern Benguela (32.3° S).

Highlights

► Latitudinally dependent diurnal wind-driven internal waves enhance vertical mixing. ► Vertical mixing drives surface layer retention and coastal warming during upwelling. ► Diurnal wind-driven oscillations of the upwelling front explain observed temperatures. ► Bottom slope steepness controls internal wave generation at the coast.

Keywords : Land-sea breeze, Inertial oscillations, Internal waves, Vertical mixing, Critical latitude, Coastal upwelling

1 Introduction

2 The land-sea breeze is a ubiquitous feature of the world's coastlines, driv-
3 ing diurnal wind variability which is detectable several hundred kilometres
4 from the coast (Gille et al., 2003, 2005). At latitudes of $\phi = 30^\circ$ N/S (re-
5 ferred to as the 'critical latitude' throughout this paper) the inertial frequency
6 ($f = 2\Omega \sin \phi$) is also diurnal, giving rise to resonance between the land-sea
7 breeze and the local inertial response of the ocean (Simpson et al., 2002;
8 Hyder et al., 2002). As the critical latitude intersects all four of the major
9 Eastern Boundary Upwelling Systems (EBUS), diurnal-inertial resonance is
10 a common feature of these systems. Indeed, many documented observations
11 of surface rotary diurnal currents in EBUS have been attributed to land-sea
12 breeze wind forcing (Hyder et al., 2011, and references therein). The physi-
13 cal and biogeochemical functioning of EBUS is however largely understood in
14 terms of the upwelling/relaxation paradigm, which responds to sub-diurnal
15 wind variability (i.e. time-scales of longer than one day). The influence of the
16 land-sea breeze on these systems is therefore typically assumed to be of low
17 importance. High amplitude diurnal-inertial currents are however known to
18 enhance shear-driven vertical mixing in coastal upwelling systems near the
19 critical latitude, as revealed from observational evidence (Aguiar-González
20 et al., 2011; Lucas et al., 2014) and numerical experiments (Fearon et al.,
21 2020). Here, we build on the 1D-vertical model experiments of Fearon et al.
22 (2020) by introducing a 2D-vertical model which includes the cross-shore di-
23 mension, allowing us to more fully explore the role of land-sea breeze forcing
24 in the context of coastal upwelling systems near the critical latitude.

25 Near-inertial rotary currents are commonly excited by surface wind vari-
26 ability, either through an impulsive wind stress such as a storm event (e.g.
27 D’Asaro et al., 1995), or through periodic forcing near the inertial frequency
28 (e.g. Simpson et al., 2002). In the case of land-sea breeze forcing near the
29 critical latitude, the surface near-inertial currents can be largely attributed
30 to the diurnal anticyclonic rotary component of the winds (τ^{ac}), as this is
31 the component of the forcing which rotates at the same frequency and in
32 the same direction as the forced surface inertial oscillations (Fearon et al.,
33 2020). Simple linearly damped slab models have been widely used to model
34 the surface mixed layer response to wind forcing, in some cases showing
35 reasonable agreement with observations of near-inertial rotary currents (e.g.
36 Pollard and Millard, 1970; Pollard, 1980; Jarosz et al., 2007). The presence
37 of a land boundary introduces a cross-shore no-flow condition at the coast
38 which drives a barotropic response via a cross-shore surface elevation gradi-
39 ent. The barotropic response simultaneously dampens the wind-driven sur-
40 face inertial oscillations and introduces subsurface inertial oscillations with
41 an opposite phase to those in the surface layer (Craig, 1989; Simpson et al.,
42 2002). Although the surface elevation gradient response behaves as an off-
43 shore propagating barotropic wave, high wave speeds ($c_0 = \sqrt{gH}$) cover
44 typical continental shelf widths in a tiny fraction of an inertial period (Chen
45 et al., 2017; Shearman, 2005), and as such can be interpreted as a locally
46 forced response. Indeed, the first order cross-shore surface elevation gradi-
47 ent, termed the ‘Craig approximation’, can be applied in 1D-vertical models
48 as a forcing term to reproduce the 180° phase shift between surface and sub-
49 surface currents (Hyder et al., 2011; Fearon et al., 2020). Throughout this
50 paper we refer to the ‘forced response’ to land-sea breeze forcing near a land
51 boundary as the superposition of the local wind-driven surface mixed layer
52 response and the opposing first order barotropic response.

53 Horizontal convergences/ divergences in the locally forced near-inertial
54 oscillations drive vertical pumping of the pycnocline, thereby initiating near-
55 inertial internal waves which can propagate away from the generation zone
56 (Alford et al., 2016). Near-inertial motions near coastlines are therefore a
57 combination of the locally forced response and the offshore propagating inter-
58 nal wave response generated by convergence/ divergence at the land bound-
59 ary (Millot and Crépon, 1981). Simple two-dimensional, linear, flat-bottom,
60 two-layer, coastal wall models have been widely used to study the inertial
61 response to an impulsive wind stress near the coast (e.g. Millot and Crépon,
62 1981; Kundu et al., 1983; Shearman, 2005; Kelly, 2019). Such models suggest

63 that the baroclinic wave generation at the land boundary plays an important
 64 role in reducing near-inertial oscillations towards the coast (Shearman, 2005).
 65 Variable bottom topography has however also been indirectly identified as a
 66 means of controlling the cross-shelf variation of near-inertial motions through
 67 its influence on the cross-shore pressure gradient (Chen and Xie, 1997). The
 68 3D numerical experiments of Zhang et al. (2010) elucidated the latitudinal
 69 dependence of near-inertial motions, internal waves and associated diapycnal
 70 mixing in response to land-sea breeze forcing at a coast. The propagation
 71 of the internal wave energy away from the coastline is dictated by the dis-
 72 persion relation for Poincare waves, which must be satisfied for propagating
 73 wave solutions to exist:

$$74 \quad \omega^2 = f^2 + c_1^2 k^2 \quad (1)$$

75 where ω is the frequency of the waves, k is the horizontal wavenumber
 76 (considering the cross-shore dimension alone) and c_1 is the phase speed of
 77 the first baroclinic mode internal wave. Zhang et al. (2010) showed that
 78 under land-sea breeze forcing, diapycnal mixing is maximised near the critical
 79 latitude of 30° N/S, where internal wave energy is trapped at the coastline
 80 due to the low group speed of Poincare waves. Diapycnal mixing is reduced at
 81 higher and lower latitudes due to the reduction of the resonance phenomenon,
 82 while at lower latitudes energy is able to propagate offshore in the form
 83 of Poincare waves and contribute to weak diapycnal mixing outside of the
 84 forcing area.

85 While the forced and internal wave responses near a land boundary can
 86 be viewed as distinct processes, their similar vertical current structures and
 87 frequencies complicate the interpretation of observations and model output.
 88 Notably, the 180° phase shift between surface and subsurface near-inertial
 89 oscillations, commonly observed near land boundaries, can be easily misin-
 90 terpreted as a true first baroclinic mode. This vertical current structure can
 91 however be reproduced in 1D models (Hyder et al., 2011; Fearon et al., 2020)
 92 or homogeneous models (Chen et al., 2017; Pettigrew, 1980) which exclude
 93 internal wave physics. The apparent first baroclinic mode vertical current
 94 structure is often cited as a source of shear-driven mixing, however the 1D
 95 model experiments of Fearon et al. (2020) showed that local shear-driven
 96 mixing is maximised in the absence of a land boundary, and that the in-
 97 troduction of the 180° phase shift between surface and subsurface layers, via
 98 the ‘Craig approximation’, serves to dampen shear production and diapycnal

99 mixing. True first baroclinic mode near-inertial internal waves are however
100 believed to play an important role in driving shear-driven turbulence and
101 diapycnal mixing (Xing et al., 2004; Zhang et al., 2010). Separating forced
102 and internal wave motions near a coastline has been achieved in simplified
103 two-layer models (Kelly, 2019), or inferred through comparisons of homoge-
104 nous vs stratified experiments using primitive equation models (Chen et al.,
105 2017). The relative contribution of the forced and internal wave responses
106 to diapycnal mixing however remains largely unstudied.

107 Within the context of coastal upwelling systems near the critical lati-
108 tude, enhanced diapycnal mixing driven by diurnal-inertial resonance has
109 been identified as an important mechanism for nutrient enrichment of the
110 surface layer, with potentially significant implications for primary productiv-
111 ity (Aguilar-González et al., 2011; Lucas et al., 2014; Fearon et al., 2020). In
112 addition to biological implications, the deepening of the surface mixed layer
113 by the land-sea breeze can also influence sub-inertial processes. For instance,
114 steepened horizontal isotherms due to land-sea breeze-driven diapycnal mix-
115 ing have been identified as playing a role in enhancing sub-inertial alongshore
116 geostrophic flows (Nam and Send, 2013). However, the implications of the
117 land-sea breeze on sub-inertial upwelling has as far as we know not been
118 explicitly studied.

119 Here, we introduce a cross-shore 2D-vertical model to study the effects of
120 land-sea forcing on coastal upwelling systems near the critical latitude. We
121 make use of the 1D-vertical model configuration of Fearon et al. (2020) as a
122 proxy for the forced response, allowing us to separate the effects of the forced
123 and internal wave responses. The 2D model is used to elucidate the effect
124 of latitude and bottom slope on the cross-shore variability in diurnal-inertial
125 currents, thermocline displacements and diapycnal mixing due to land-sea
126 breeze forcing near the critical latitude. Analytically configured experiments
127 for this purpose are set up in an analogous way to the 1D experiments de-
128 scribed in Fearon et al. (2020), and we employ the 1D model solution as
129 the offshore open boundary condition for the 2D model. The quantification
130 of diapycnal mixing is aided by initialising subsurface waters with a passive
131 tracer. We then further explore the implications of the land-sea in the con-
132 text of coastal upwelling systems by considering experiments forced by the
133 land-sea breeze alone, upwelling winds alone, and a combination of the two.
134 We initialise subsurface waters of these experiments with lagrangian floats
135 to track how subsurface waters are modified through land-sea breeze-driven
136 diapycnal mixing. As in Fearon et al. (2020), we again compare a realisti-

137 cally configured model to the observations of Lucas et al. (2014), located in
138 St Helena Bay in the Southern Benguela Upwelling System (Figure 1).

139 2. Methods

140 2.1. *In-situ observations*

141 We make use of horizontal current velocity and temperature observa-
142 tions from a mooring array described in Lucas et al. (2014), located in St
143 Helena Bay in the Southern Benguela Upwelling System (Figure 1). The
144 three fixed moorings are aligned roughly perpendicular to the local coast-
145 line, making them particularly relevant for identifying the cross-shore pro-
146 cesses of interest in this paper. Horizontal current velocity data are ob-
147 tained from bottom-mounted Acoustic Doppler Current Profilers (ADCP),
148 while temperature data are obtained from Wirewalker wave-powered pro-
149 filers (Rainville and Pinkel, 2001; Pinkel et al., 2011). Velocity and tem-
150 perature data are available at a temporal frequency of 10 min and at ver-
151 tical resolutions of 1 m and 0.25 m, respectively. All observations are fil-
152 tered in time to provide a two hour running mean at 30 min intervals,
153 sufficient for revealing processes at the diurnal-inertial frequency of inter-
154 est. As per Fearon et al. (2020), we revisit only the 7-14 March 2011
155 upwelling event, having been identified as a period which clearly demon-
156 strates the response of a highly stratified two layer system to a combination
157 of upwelling favourable winds and strong diurnal wind variability (Lucas
158 et al., 2014). The in-situ observations are available for download via the fol-
159 lowing Digital Object Identifiers (DOIs): Wirewalker data (<https://doi.org/10.15493/dea.mims.26052100>), ADCP data (<https://doi.org/10.15493/dea.mims.26052101>).

162 2.2. *Ocean model*

163 The ocean model employed in this study is the Coastal and Regional
164 Ocean COmmunity model (CROCO) (<http://www.croco-ocean.org/>), an
165 ocean modelling system built upon ROMS_AGRIF (Shchepetkin and McWilliams,
166 2005). CROCO is a free-surface, terrain-following coordinate oceanic model
167 which solves the Navier-Stokes primitive equations by following the Boussi-
168 nesq and hydrostatic approximations.

169 1D model experiments employ the standalone 1D version of the code
170 described in Fearon et al. (2020), in which we retain the coast-normal hori-
171 zontal pressure gradient as a forcing term computed from the ‘Craig approx-

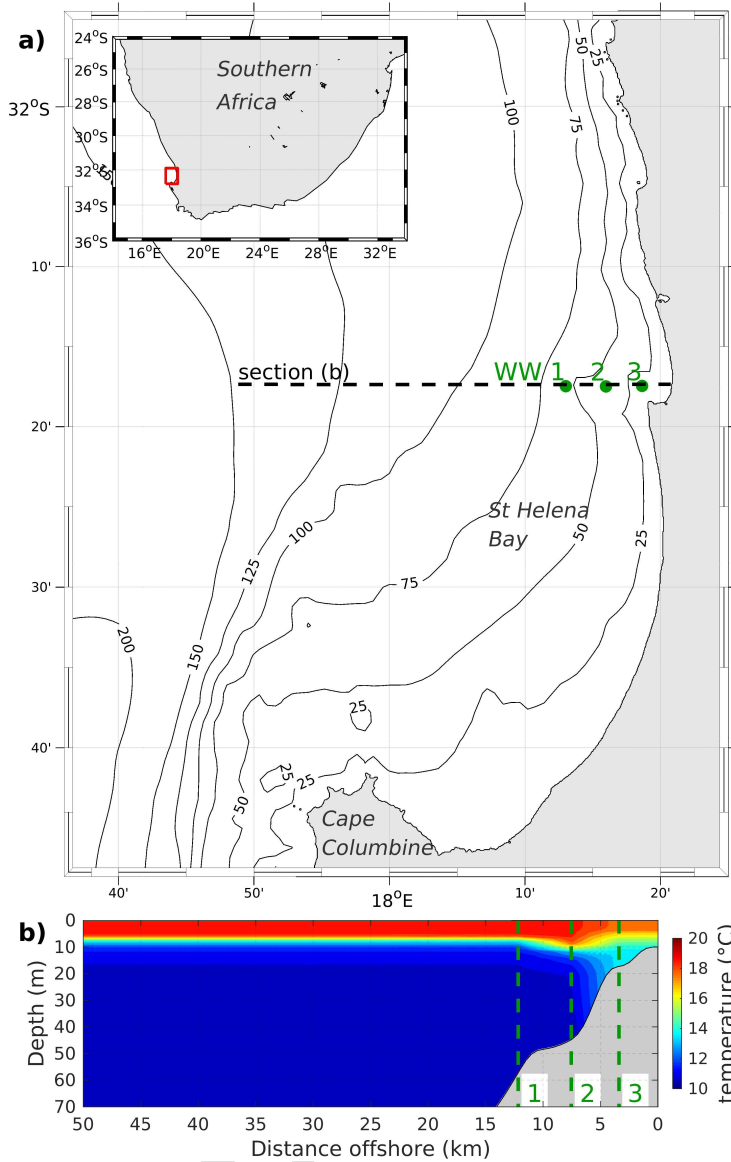


Figure 1: (a) Locality map for the Wirewalker (WW) cross-shore mooring array of Lucas et al. (2014). The black dotted line denotes the location of the section shown in (b). (b) Temperature initial condition for the realistically configured 2D model, interpolated and extrapolated from the observations on 7 March 2011. Only the closest 50 km to land boundary are shown. Bathymetry is interpolated from digital navigational charts for the region provided by the Hydrographer of the SA Navy.

172 imation'. The 1D model code can be downloaded via the following DOI:
 173 <https://doi.org/10.15493/dea.mims.26052102>.

174 2D model experiments are configured using the V1.0 official release of
 175 the 3D CROCO code. This is achieved by employing a shore-perpendicular
 176 grid with only 5 grid cells in the alongshore dimension (Section 2.2.1), main-
 177 taining constant alongshore bathymetry (Section 2.2.3) and making use of
 178 periodic boundary conditions along the cross-shore boundaries of the model
 179 (Section 2.2.4). Under these conditions, alongshore gradients are essentially
 180 zero and continuity dictates that divergence/ convergence of the cross-shore
 181 flow ($\frac{\partial u}{\partial x}$) must be compensated by vertical motion ($\frac{\partial w}{\partial z}$).

182 The parameterisation of vertical eddy viscosity and diffusivity are carried
 183 out in the present study in accordance with the k - ε turbulent closure scheme
 184 within the Generic Length Scale (GLS) formulation (Umlauf and Burchard,
 185 2003, 2005). Implementation of the scheme within CROCO, and default set-
 186 tings adopted in this study, are provided in Appendix A of Fearon et al.
 187 (2020). We use an upstream-biased, dissipative horizontal advection scheme
 188 for momentum, while horizontal advection of tracers is discretized using
 189 a split and rotated third-order upstream-biased numerical scheme (March-
 190 esiello et al., 2009). No explicit lateral viscosity is added in the model, except
 191 in sponge layers at the western open lateral boundary. As described in Fearon
 192 et al. (2020), bottom friction is parameterised using a quadratic drag law,
 193 where the bottom roughness length parameter is taken as 0.1 m. A nonlinear
 194 equation of state adapted from Jackett and McDougall (1995) is used for the
 195 computation of density.

196 Both analytically and realistically configured experiments are employed
 197 in this paper, as summarised in Table 1. The analytically configured experi-
 198 ments (row ID's 1 and 2 of Table 1) are designed to elucidate the processes
 199 of interest and to carry out sensitivity tests to relevant variables, while the
 200 realistically configured simulation (row ID 3 of Table 1) is compared with
 201 observations from the cross-shore mooring array of Lucas et al. (2014) over
 202 the 7-14 March 2011 upwelling event. For consistency, the 1D model con-
 203 figurations presented here are identical to those presented in Fearon et al.
 204 (2020), while the 2D model settings remain consistent with the 1D model
 205 settings where possible, as described below.

206 *2.2.1. Spatial and temporal discretisation*

207 All 2D experiments use a model domain with horizontal dimensions of
 208 100 km (E-W) \times 2.5 km (N-S), which is discretised using a regular grid of

Table 1: Summary of model experiments

ID	Description	Surface forcing	Bathymetry	Latitude
1	Diurnal-inertial resonance at a land boundary (Section 3.1)	$\tau^{ac0} = 0.03 \text{ N m}^{-2}$, $\bar{\tau}_y = 0 \text{ N m}^{-2}$	flat bottom	20°S, 30°S, 40°S
			1:200, 1:500	30°S
2	Diurnal-inertial resonance in the presence of upwelling (Section 3.3)	$\tau^{ac0} = 0 \text{ or } 0.03 \text{ N m}^{-2}$, $\bar{\tau}_y = 0 \text{ or } 0.05 \text{ N m}^{-2}$	1:200	30°S
3	Case study of St Helena Bay (Section 3.4)	CSAG WRF model	interpolated from navigation charts	32.3°S

209 500 m resolution (i.e. a 200×5 grid). We use 100 sigma layers to discretise
 210 the vertical dimension. A baroclinic time-step of 40 s is used to integrate
 211 the model solution over a period of 7 days from initialisation, typical of the
 212 time-scale of upwelling events. 40 barotropic time-steps are computed within
 213 each baroclinic time-step. Instantaneous model output at 30 min intervals
 214 is filtered in time to provide a two hour running mean at each time-step,
 215 consistent with the processing of observations.

216 2.2.2. Surface forcing

217 In the case of analytically configured experiments (row ID's 1 and 2 of
 218 Table 1) the land-sea breeze is approximated as a diurnal anticyclonic (anti-
 219 clockwise in the southern hemisphere) rotating wind stress (τ^{ac}) of constant
 220 amplitude (τ^{ac0}), which is applied as a spatially constant forcing. Hyder
 221 et al. (2011) report diurnal anticyclonic rotary wind stress amplitudes from
 222 in-situ observations along the Namibian coastline of between 0.004 N m^{-2} and
 223 0.051 N m^{-2} , depending on the location and observation period. Analysis of
 224 surface wind stress derived from a 3 km horizontal resolution atmospheric
 225 model indicates large spatial variability in the diurnal rotary component of
 226 winds over the Southern Benguela, with a notable enhancement over St He-
 227 lena Bay (Fearon et al., 2020). In this region, the mean amplitude of the
 228 diurnal anticyclonic rotary component of the wind stress, as derived from
 229 7 day windows over upwelling favourable months, was found to be around
 230 0.03 N m^{-2} . We therefore adopt a value of $\tau^{ac0} = 0.03 \text{ N m}^{-2}$ for all ana-
 231 lytically configured experiments presented in this paper. Time-series of the
 232 analytically configured rotary surface wind stress components ($\tau^{ac} = (\tau_x^{ac},$
 233 $\tau_y^{ac})$) are shown in Figure 2a.

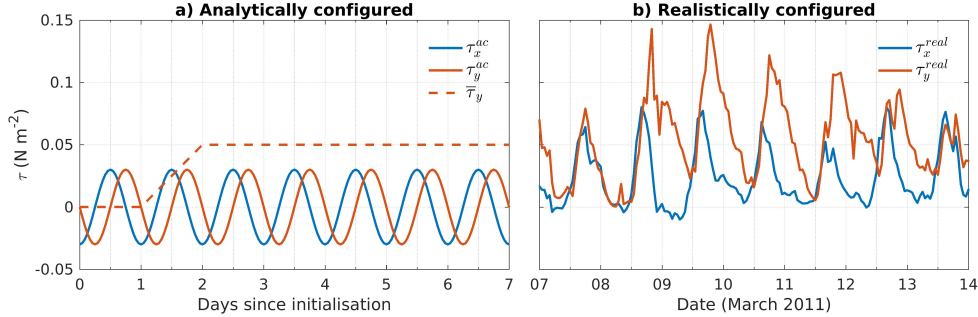


Figure 2: Time-series of surface wind stress components applied in the model configurations. (a) $\tau^{ac} = (\tau_x^{ac}, \tau_y^{ac})$ represents a diurnal anticlockwise (anticyclonic in the southern hemisphere) rotating wind stress with a constant amplitude (τ^{ac0}) of 0.03 N m^{-2} , used to represent land-sea breeze forcing. $\bar{\tau}_y$ is used to represent an upwelling favourable wind stress in the southern hemisphere. (b) Wind stress components derived from the 3 km resolution CSAG WRF atmospheric model at the outer mooring (WW1 in Figure 1) over the duration of the simulated 7-14 March 2011 upwelling event.

234 When upwelling is included in these experiments (row ID 2 of Table 1),
 235 it is simulated through the application of a mean alongshore wind stress
 236 ($\bar{\tau}_y$) of 0.05 N m^{-2} , being typical of the 7 day mean alongshore wind stress
 237 for St Helena Bay over upwelling favourable months (Fearon et al., 2020).
 238 $\bar{\tau}_y$ is linearly ramped up from zero to 0.05 N m^{-2} over the second day of
 239 the simulation, allowing for the preconditioning of the water column by the
 240 land-sea breeze over the first two days before the full impact of the mean
 241 alongshore wind stress is felt by the surface Ekman boundary layer. Time-
 242 series of the analytically configured surface wind stress components are shown
 243 in Figure 2a. Surface heat fluxes are excluded in analytically configured
 244 experiments.

245 In the case of the realistically configured experiment (row ID 3 of Ta-
 246 ble 1), bulk parameterisation is adopted for the computation of surface wind
 247 stress and surface net heat fluxes (Fairall et al., 1996, 2003) using hourly
 248 atmospheric model output from a 3 km resolution Weather Research and
 249 Forecasting (WRF) model configuration developed by the Climate Systems
 250 Analysis Group (CSAG) at the University of Cape Town (UCT). The atmo-
 251 spheric simulation forms part of the Wind Atlas for South Africa (WASA)
 252 project (<http://www.wasaproject.info/>) and has been validated against a
 253 number of land-based weather stations, including one deployed at the south-
 254 ern end of St Helena Bay over a three year period (Lennard et al., 2015).

255 Figure 2b presents time-series of surface wind stress components at the outer
 256 mooring (WW1 in Figure 1) over the duration of the simulated 7-14 March
 257 2011 upwelling event.

258 *2.2.3. Bathymetry and latitude*

259 Flat bottom analytically configured experiments employ a constant water
 260 depth of 100 m depth, and we test the sensitivity of the model solution to
 261 cross-shore bottom slopes of 1:200 and 1:500 (row ID's 1 and 2 of Table 1).
 262 Flat bottom experiments are considered to be a useful reference against pre-
 263 vious flat bottom model studies, where internal wave generation at the land
 264 boundary controls the onshore decrease in horizontal near-inertial current
 265 velocities (e.g. Shearman, 2005). Different bottom slopes allow us to test
 266 the role of the cross-shore pressure gradient (which increases with decreas-
 267 ing water depth according to the 'Craig approximation') in controlling the
 268 cross-shelf decrease in near-inertial motions. This mechanism was found to
 269 be a dominant driver of cross-shelf variability of near-inertial motions in the
 270 realistically configured 2D model of Chen and Xie (1997).

271 A maximum offshore water depth of 100 m is used for all experiments,
 272 while a minimum water depth of 20 m is employed at the land boundary.
 273 We maintain a constant alongshore bathymetry in all experiments. The
 274 realistically configured experiment (row ID 3 of Table 1) employs a nearshore
 275 bathymetry (i.e. shallower than 100 m) which is interpolated from digital
 276 versions of the most detailed available navigation charts for the region, as
 277 provided by the Hydrographer of the South African Navy (Figure 1).

278 Baseline experiments adopt a latitude of 30°S for testing the pure case
 279 of diurnal-inertial resonance, however we also test the model sensitivity to
 280 latitudes of 20°S and 40°S (row ID's 1 and 2 of Table 1). Latitude is expected
 281 to play a dominant role in land-sea breeze-driven near-inertial motions both
 282 through local resonance effects (Simpson et al., 2002; Hyder et al., 2002;
 283 Fearon et al., 2020) and internal wave generation Zhang et al. (2010). The
 284 realistically configured experiment adopts the realistic cross-shore mooring
 285 array latitude of $\sim 32.3^\circ\text{S}$ (row ID 3 of Table 1).

286 *2.2.4. Boundary conditions*

287 The northern and southern boundaries are defined as periodic boundary
 288 conditions, such that all the outflows (inflows) at the southern boundary
 289 are inflows (outflows) at the northern boundary. Given the constant along-
 290 shore bathymetry, the model is effectively a 2D configuration. The eastern

291 boundary is defined as a closed land boundary, while the western boundary
292 is applied as an open boundary condition. The open boundary requires the
293 prescription of temperature (T), salinity (S) and velocity components (u, v).
294 Additionally, we prescribe a passive tracer concentration boundary condition
295 (C , in arbitrary tracer units per volume, ATU m^{-3}), used to aid the quantifi-
296 cation of diapycnal mixing (see Section 2.3). Relaxation times of 1/4 day and
297 1 day are adopted for inward and outward radiation, respectively, implying
298 strong relaxation to the specified boundary values. A sponge layer of 10 km
299 (20 grid cells) is used to gradually ramp up the model solution in the interior
300 of the domain to the applied boundary values within the sponge layer. The
301 prescribed boundary conditions over each experiment are obtained from the
302 solution of an analogously configured 1D model, integrated over the simu-
303 lation period using the same initial condition and forcing as applied at the
304 offshore extent of the 2D model.

305 *2.2.5. Initial conditions*

306 We remain consistent with the 1D model experiments of Fearon et al.
307 (2020) in defining initial conditions for the 2D model experiments. Simula-
308 tions are initialised from rest using a constant salinity of 35 and a temper-
309 ature profile defined either analytically or from observations. In the case of
310 analytically configured experiments (row ID's 1 and 2 of Table 1), the initial
311 vertical profile for temperature is defined using a hyperbolic tangent function
312 which effectively creates a two layer system with 16°C surface water over-
313 laying 10°C subsurface water and a maximum stratification located at 10 m
314 depth (see Fearon et al. (2020) for details). The realistic model configuration
315 (row ID 3 of Table 1) is initialised using measured temperature profiles from
316 the three moorings of Lucas et al. (2014) at the start of the simulation (7
317 March 2011). Linear interpolation is used to define the initial temperature
318 in between the offshore and inshore moorings, while temperature data from
319 the offshore (inshore) mooring is used to extrapolate offshore (inshore) of the
320 moorings. While it is acknowledged that the temperature section will not be
321 realistic at the offshore extent of the model, the intention is only to provide
322 a stable offshore boundary so that the nearshore processes in the vicinity
323 of the observations can be assessed. The resulting initial temperature sec-
324 tion is shown in Figure 1b. Experiments are initialised with passive tracer
325 concentrations of 1 and 0 ATU m^{-3} in the subsurface and surface layers,
326 respectively.

327 *2.3. Data analysis*

328 As alongshore variability in the model is negligible, we only present plots
329 and analysis of the 2D model output along a cross-shore section corresponding
330 to the centre (i.e. third) alongshore grid cell.

331 We remain consistent in the computation of diapycnal mixing diagnostics
332 as described in Fearon et al. (2020). The thermocline depth (H_s) is defined
333 as the depth of a given isotherm, being the 11° C isotherm in the case of ana-
334 lytically configured experiments, and the 12.5° C isotherm in the case of the
335 realistically configured experiment. The cumulative diapycnal mixing over
336 each experiment is quantified through the diagnostic variable C_s , computed
337 by integrating the passive tracer (C) over the surface layer of the model (i.e.
338 from H_s to the surface). The depth averaged velocity vector over the surface
339 layer of the model is used to indicate the surface layer current response in
340 the model ($\vec{u}_s = (u_s, v_s)$). The analysis of diapycnal mixing is further aided
341 by extracting the vertical turbulent diffusivity (K_{Tv}), as output from the k - ϵ
342 turbulent closure scheme of the model. K_{Tv} is extracted at the thermocline
343 depth as an indication of interfacial mixing in the model.

344 We further use the vertical displacement of the thermocline in order to
345 diagnose internal wave generation and propagation in the model. To this
346 end, the thermocline depth is presented as the displacement from the daily
347 running average thermocline depth, allowing us to isolate the super-diurnal
348 variability in the thermocline depth induced by the applied land-sea breeze
349 forcing. Thermocline displacements are compared with the theoretical inter-
350 nal Rossby radius of deformation (R_d):

$$351 \quad R_d = \frac{c_1}{f} \quad (2)$$

352 and further compared with the theoretical celerity of a long wave propa-
353 gating at the interface of a two layer system:

$$354 \quad c_1 = \sqrt{g' \frac{H_s H_b}{H_s + H_b}} \quad (3)$$

355 where $g' = \frac{\Delta\rho}{\rho} g$ is the reduced gravity acceleration. H_s and H_b are the
356 thicknesses of the surface and bottom layers, respectively, as defined by the
357 depth of the thermocline. g' is estimated from the temperature difference
358 between the surface and subsurface layers and a linear expansion coefficient
359 $\alpha = 2 \times 10^{-4} \text{ K}^{-1}$.

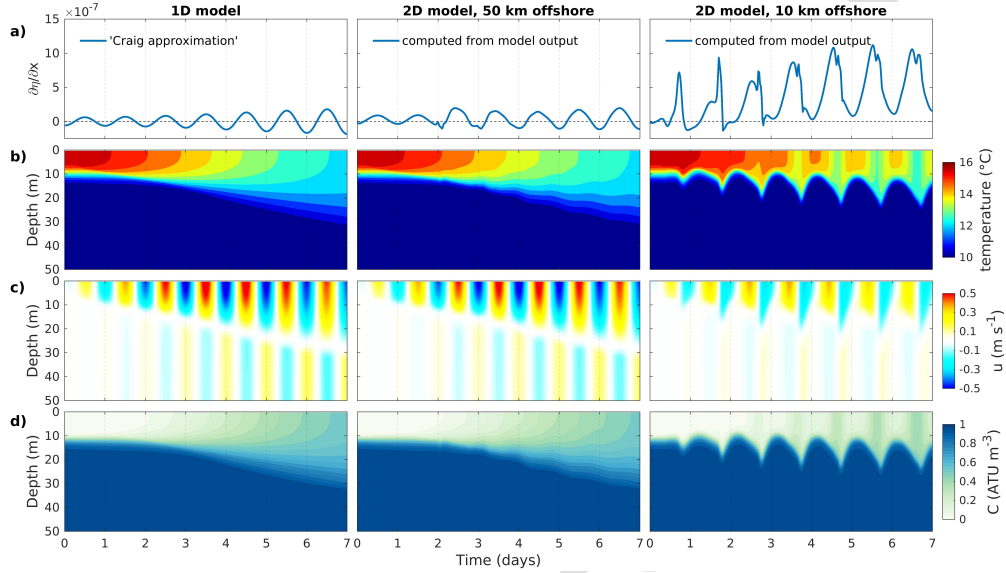


Figure 3: Flat bottom case, water depth = 100 m. Comparison of the 1D model and the 2D model at distances of 50 km and 10 km from the land boundary. Time-series of: (a) surface elevation gradient ($\frac{\partial \eta}{\partial x}$), represented by the ‘Craig approximation’ in the case of the 1D model, and computed from the model output in the case of the 2D model; (b) vertical profile of temperature; (c) vertical profile of cross-shore velocity (u); (d) vertical profile of passive tracer concentration (C). Results are computed from a 7 day integration of the models with input parameters $\tau^{ac0} = 0.03 \text{ N m}^{-2}$, latitude = 30° S , initial stratification (ΔT) = 6° C , initial depth of maximum stratification = 10 m.

360 When an upwelling favourable wind is included in the model (row ID’s 2
 361 and 3 of Table 1), the location of the upwelling front is computed from the
 362 outcropping of the diagnostic isotherm used to define the thermocline depth.

363 3. Results

364 3.1. Diurnal-inertial resonance at a land boundary

365 We begin by considering the response of a flat bottom 100 m deep ocean
 366 consisting of an analytically derived two-layer water column to a constant
 367 amplitude diurnal anticlockwise rotating wind stress at 30° S (the pure case
 368 of diurnal-inertial resonance). The amplitude of the applied wind stress is
 369 0.03 N m^{-2} , and we exclude any sub-inertial upwelling from a mean along-
 370 shore wind stress. Figure 3 presents the evolution of the model output for
 371 both the 1D model configuration, as presented in Fearon et al. (2020), and

372 the 2D model configuration introduced in this paper at distances of 50 km
 373 and 10 km from the land boundary. Figure 4 presents Hovmöller diagrams for
 374 the 2D experiment showing cross-shore variability in thermocline displace-
 375 ments, $\log_{10} K_{Tv}$ at the depth of the 11° C isotherm (taken to represent the
 376 thermocline), and the passive tracer concentration integrated from the 11° C
 377 isotherm to the surface (C_s).

378 The results indicate that the 1D and 2D model solutions are almost iden-
 379 tical at a distance of 50 km offshore. In the case of the 1D model, the cross-
 380 shore surface elevation gradient ($\frac{\partial \eta}{\partial x}$) is an applied forcing term computed
 381 from the ‘Craig approximation’ for the barotropic response imposed by the
 382 no-flow condition perpendicular to the land boundary, while this response is
 383 generated dynamically in the 2D model. The near-perfect agreement between
 384 the two models indicates that the reduced physics of the 1D model provides
 385 a sufficient description of the processes at this location, and therefore the
 386 enhanced shear-driven vertical mixing here is driven by the locally forced re-
 387 sponse alone. These processes are explored in detail in Fearon et al. (2020),
 388 and are therefore not elaborated further here. The excellent agreement also
 389 provides support for the methodology of using the 1D model solution as the
 390 offshore boundary condition for the 2D model. Indeed, this approach largely
 391 prevented spurious internal wave generation at the open boundary.

392 At a distance of 10 km offshore, the 2D model solution deviates signifi-
 393 cantly from the 1D solution, as the linear assumptions of the ‘Craig approxi-
 394 mation’ are invalid. Here, in close proximity to the land boundary, horizontal
 395 gradients in cross-shore currents are high and the non-linear advection terms
 396 become important. Onshore (offshore) surface currents lead to convergence
 397 (divergence) and the downward (upward) displacement of the thermocline.
 398 The forced thermocline displacements have a diurnal periodicity, consistent
 399 with the diurnal periodicity of the wind forcing. Unlike at 50 km offshore,
 400 the results at 10 km offshore reveal diurnal variability in surface tempera-
 401 ture and surface tracer concentrations (Figure 3b), reflecting the advection of
 402 horizontal gradients which are set up by high spatial variability in diapycnal
 403 mixing. The amplitude of the surface inertial oscillations are significantly
 404 weaker at 10 km offshore than at 50 km offshore (Figure 3c), implying re-
 405 duced vertical shear and therefore reduced diapycnal mixing, as reflected in
 406 the surface tracer concentration (Figure 3d). We note that although the
 407 alongshore current velocity data are not plotted, the oscillatory nature of the
 408 currents dictates a very similar vertical structure in the alongshore currents,
 409 the only notable difference being a 6 hour phase shift when compared to the

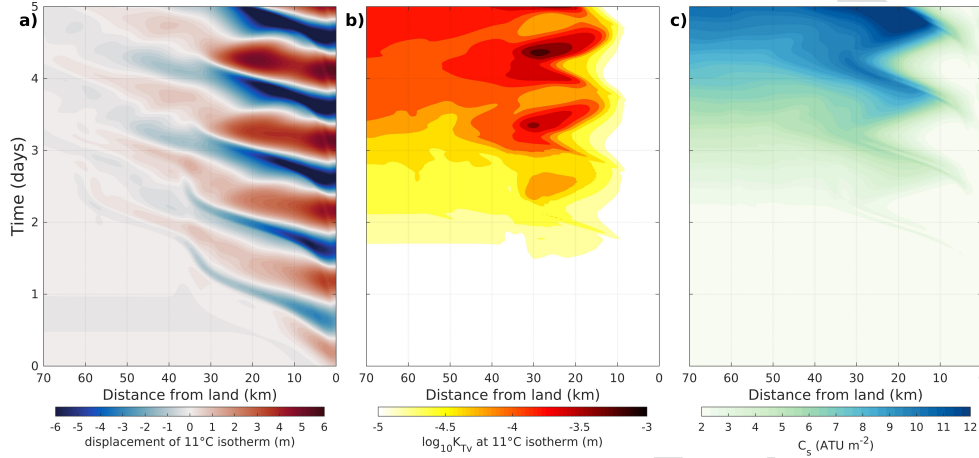


Figure 4: Hovmöller diagrams for the 2D experiment shown in Figure 3. (a) Vertical displacement of the 11° C isotherm from the daily running average isotherm depth (positive is upward); (b) $\log_{10} K_{T_v}$ at the 11° C isotherm; (c) passive tracer concentration integrated from the 11° C isotherm to the surface (C_s).

410 cross-shore currents.

411 Hovmöller diagrams of the 2D experiment (Figure 4) reveal that the vertical
 412 displacements of the thermocline propagate offshore, but are largely
 413 absent 50 km from the land boundary, indicating an evanescent internal
 414 wave response. Between 10 km and 50 km offshore, diapycnal mixing is
 415 shown to respond to a combination of the forced response and the internal
 416 wave response, generating diurnal peaks of enhanced mixing at the diagnostic
 417 thermocline (Figure 4b). The net result is an enhancement of the cumulative
 418 diapycnal mixing over and above that induced by the forced response alone,
 419 as revealed by a locally enhanced passive tracer concentration in the surface
 420 layer of the model (Figure 4c).

421 3.2. Sensitivity tests to latitude and bottom slope

422 The evanescent nature of the internal wave response to land-sea breeze
 423 forcing is governed by the latitude, as previously explored by Zhang et al.
 424 (2010), and revealed in Figure 5. Here, we contrast the simulation presented
 425 in Figures 3 and 4 (latitude = 30°S) against identical simulations run at
 426 latitudes of 20°S and 40°S. The results are presented as Hovmöller diagrams
 427 of thermocline displacement, as well as the amplitude of the surface layer
 428 velocity ($|\vec{u}_s|$) and the passive tracer concentration integrated over the surface

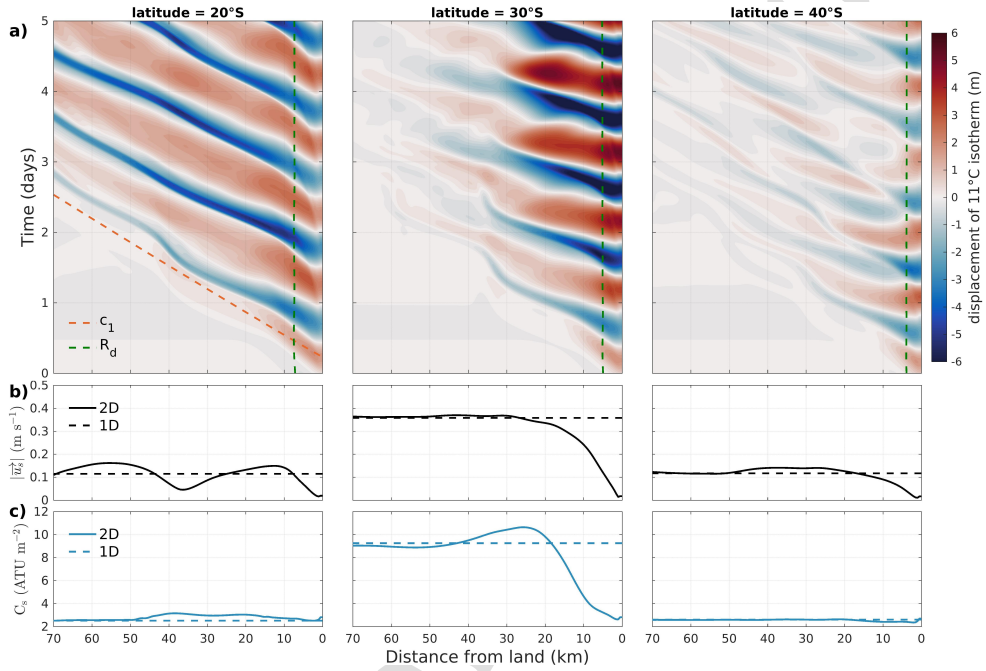


Figure 5: Flat bottom case, water depth = 100 m. Effect of latitude on cross-shore variability in thermocline displacement, current amplitude and diapycnal mixing. (a) Hovmöller diagram of the vertical displacement of the 11° C isotherm from the daily running average isotherm depth (positive is upward). The orange and green dotted lines denote the theoretical internal wave speed (c_1) and internal Rossby radius of deformation (R_d), respectively. (b) Amplitude of the surface layer velocity ($|\overline{u_s}|$), averaged over the fifth day of the simulation. (c) Passive tracer concentration integrated over the surface layer (C_s), averaged over the fifth day of the simulation. The dotted line in (b) and (c) denotes the 1D solution at the corresponding depth. Results are computed from a 7 day integration of the models with input parameters $\tau^{ac0} = 0.03 \text{ N m}^{-2}$, initial stratification ($\Delta T = 6^\circ \text{ C}$, initial depth of maximum stratification = 10 m.

429 layer (C_s), both averaged over the fifth day of each simulation. The 1D model
 430 output for $|\vec{u}_s|$ and C_s is also shown for comparative purposes, allowing us
 431 to distinguish the relative influence of the forced response (as approximated
 432 by the 1D model) and the internal waves. For consistency, $|\vec{u}_s|$ and C_s are
 433 the same metrics used to summarise the 1D model sensitivity experiments
 434 presented in Fearon et al. (2020).

435 The forced diurnal displacement of the thermocline is shown to occur
 436 landward of the internal Rossby radius of deformation (R_d , Equation 2),
 437 while the offshore propagation of the forced displacement is latitudinally de-
 438 pendent. At 20°S, the diurnal pumping of the thermocline results in offshore
 439 propagating internal waves, however these waves are evanescent at latitudes
 440 of 30°S and 40°S. Equation 1 dictates that waves can only freely propagate
 441 at latitudes where the inertial frequency f is less than the frequency of the
 442 forced diurnal displacements (i.e. equatorward of 30° N/S, such that $f < \omega$).
 443 The results indicate that the initial propagation of the internal wave away
 444 from the land boundary can be reasonably approximated by the theoretical
 445 celerity of a long wave propagating at the interface of a two layer system (c_1 ,
 446 Equation 3).

447 The amplitude of the thermocline displacement is greatest at the critical
 448 latitude of 30°S, where the amplitude of the forced surface currents and there-
 449 fore convergence/ divergence at the land boundary is greatest (Figure 5b).
 450 At this latitude, the amplitude of the surface current is shown to drop off
 451 rapidly within ~ 20 km of the land boundary, in good agreement with the
 452 flat bottom numerical experiments of Chen et al. (2017). Diapycnal mixing,
 453 as quantified through the passive tracer concentration integrated over the
 454 surface layer (C_s), is shown to be highest at a distance of ~ 25 km offshore,
 455 and is $\sim 20\%$ greater than the offshore value. As already described, the el-
 456 evated mixing is attributed to the effect of the evanescent internal waves
 457 which contribute to the mixing induced by the forced response.

458 Both surface current amplitude and diapycnal mixing are predictably
 459 lower at latitudes of 20°S and 40°S when compared to 30°S. For the 20°S
 460 experiment the signature of the offshore propagating internal waves is evident
 461 both through deviations of the surface current amplitude (Figure 5b) and
 462 elevated diapycnal mixing (Figure 5c) when compared to the forced response.
 463 This is not seen in the 40°S experiment, where the effect of the internal wave
 464 response is negligible.

465 We now add an element of realism to the experiments by contrasting the
 466 100 m water depth flat bottom configuration at 30°S already presented with

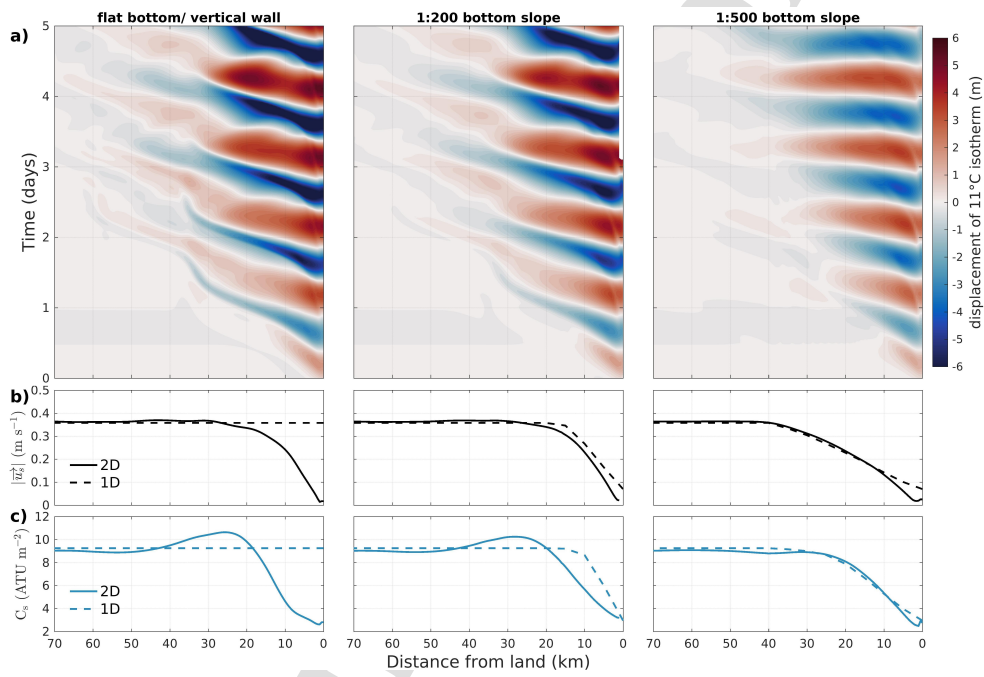


Figure 6: As per Figure 5, but testing the effect of bottom slope at a latitude of 30°S. The left panels are identical to the centre panels in Figure 5.

467 two experiments in which the water depth gradually increases away from the
 468 land boundary, using constant bottom slopes of 1:200 and 1:500, respectively
 469 (Figure 6). By way of comparison, the average bottom slope across the St
 470 Helena Bay mooring array is approximately 1:200 (Figure 1). The dashed
 471 lines in Figure 6b,c are taken from individual 1D model solutions run at 5 km
 472 increments from the land boundary, using the corresponding water depth of
 473 the 2D model. These 1D experiments provide insight into how a gradually
 474 reducing water depth impacts the forced response alone.

475 As explored in Fearon et al. (2020), the first order surface elevation gra-
 476 dient response serves to reduce surface oscillations and diapycnal mixing,
 477 providing a mechanism for gradually dampening the forced diurnal-inertial
 478 oscillations in shallower water depths toward the land boundary. The damp-
 479 ened forced response is shown to lead to a reduction in the convergence/
 480 divergence of the forced oscillations in the surface layer, and therefore a re-
 481 duction in thermocline pumping (Figure 6a). Considering the case of the
 482 1:500 bottom slope, the 2D model results for both $|\vec{u}_s|$ and C_s are almost
 483 identical to the 1D model (right panels of Figure 6b,c), indicating that the
 484 internal wave effects on currents and diapycnal mixing are negligible. As the
 485 slope becomes steeper, so the convergence/ divergence in the forced response
 486 becomes greater and the effects of the internal waves become apparent. In
 487 these cases the 1D model over-predicts both the amplitude of the surface
 488 oscillation and diapycnal mixing within 20 km of the land boundary.

489 3.3. Diurnal-inertial resonance in the presence of upwelling

490 In the case of upwelling systems, the processes associated with diurnal-
 491 inertial resonance are embedded within Ekman dynamics driven by sub-
 492 inertial wind variability. The 2D model allows us to explore the interaction
 493 of these processes by comparing simulations forced with a land-sea breeze
 494 alone ($\tau^{ac0} = 0.03 \text{ N m}^{-2}$), an upwelling wind alone ($\bar{\tau}_y = 0.05 \text{ N m}^{-2}$), and a
 495 combination of the two. The values of τ^{ac0} and $\bar{\tau}_y$ are specifically chosen so
 496 that a combination of the two is representative of a ‘typical’ upwelling event
 497 within St Helena Bay (Section 2.2.2). We adopt a bottom slope of 1:200 for
 498 these experiments, being representative of the bottom slope within St Helena
 499 Bay (Figure 1). Figure 7 presents the temporal evolution of the modelled
 500 temperature, cross-shore currents and passive tracer for each experiment at
 501 a distance of 10 km from the land boundary.

502 The processes associated with the simulation forced by the land-sea breeze
 503 alone have already been shown (Figure 3), whereby diurnal variability in

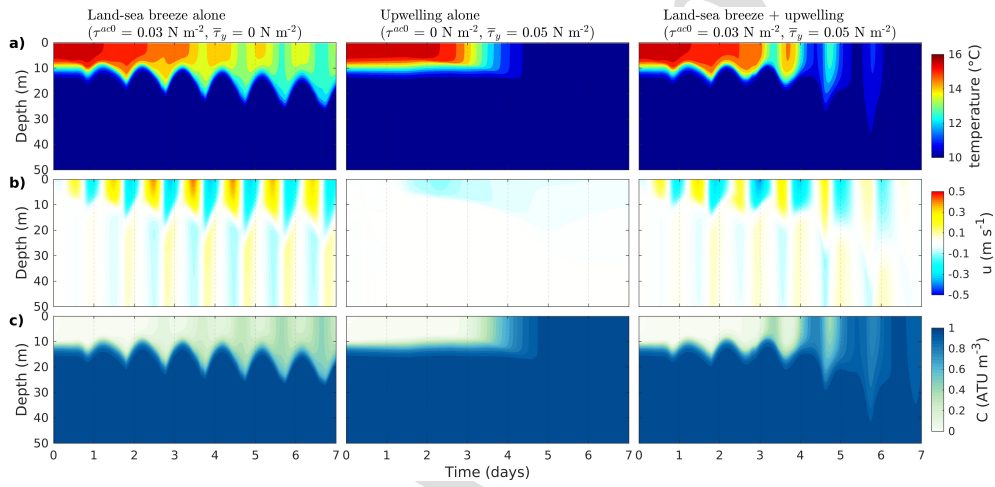


Figure 7: Effect of the land-sea breeze over a ‘typical’ upwelling event at a distance of 10 km from the land boundary. Simulations are forced with a land-sea breeze alone (left), an upwelling wind alone (middle), and a combination of the two (right). (a) Vertical profile of temperature; (b) vertical profile of cross-shore velocity (u); (c) vertical profile of passive tracer concentration (C). Results are computed from a 7 day integration of the 2D model with input parameters latitude = 30° S, bottom slope = 1:200, initial stratification (ΔT) = 6° C, initial depth of maximum stratification = 10 m.

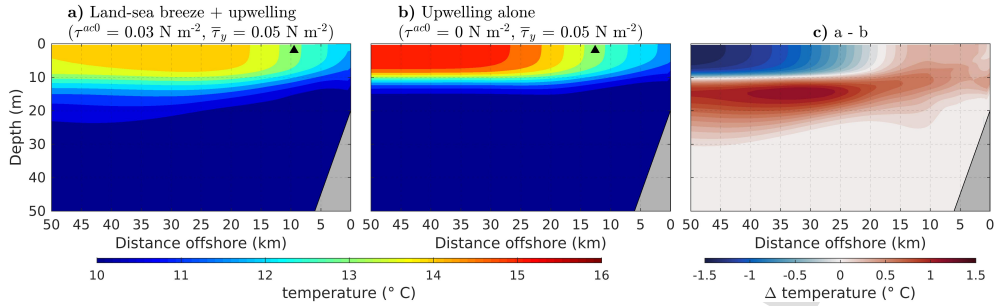


Figure 8: Modelled 7 day mean cross-shore temperature over a ‘typical’ upwelling event. The black triangles at the surface denote the mean location of the upwelling front, as computed from the outcropping of the 11° C isotherm. (a) Simulation forced with a combination of land-sea breeze and upwelling winds; (b) simulation forced with an upwelling wind alone; (c) difference between (a) and (b). Results are computed from a 7 day integration of the models with input parameters latitude = 30° S, bottom slope = 1:200, initial stratification (ΔT) = 6° C, initial depth of maximum stratification = 10 m.

504 surface temperature is driven by a combination of high spatial variability
 505 in diapycnal mixing and advection due to the surface inertial oscillations.
 506 The simulation forced by the upwelling wind alone shows considerably re-
 507 duced diapycnal mixing, and shows the offshore transport of the surface
 508 layer in response to the sustained alongshore wind stress (centre panel of
 509 Figure 7b). The combined effect of land-sea breeze and upwelling winds gen-
 510 erates super-diurnal advection of the upwelling front, leading to pronounced
 511 diurnal variability in surface temperatures. The inertial oscillations (left
 512 panels of Figure 7) are superimposed onto the sub-inertial transport (cen-
 513 tre panels of Figure 7), resulting in the repeated cross-shore back and forth
 514 advection of strong horizontal temperature gradients past the shown output
 515 location (right panels of Figure 7). Large diurnal variability in temperature
 516 will persist provided the location of interest is within one oscillation radius
 517 of the steep horizontal temperature gradients associated with the upwelling
 518 front (an oscillation velocity of 0.5 m s^{-1} at 30° S has an oscillation radius ($\frac{u}{f}$)
 519 of 6.9 km). This phenomenon is invoked in the interpretation of the results
 520 of the realistically configured experiment, as described in Section 3.4.

521 In addition to the super-diurnal advection of the upwelling front, the
 522 inclusion of the land-sea breeze is shown to significantly influence the mean
 523 cross-shore temperature over the modelled upwelling event (Figure 8). The
 524 mean location of the upwelling front is also shown, as computed from the

525 outcropping of the 11° C isotherm. At the offshore extent of the shown
 526 section (i.e. outside the influence of the upwelling front), the applied diurnal
 527 wind variability is shown to result in a significant cooling of the surface,
 528 which overlays a commensurate warming. Despite the surface cooling due to
 529 diapycnal mixing, the results indicate a net warming of mean surface waters
 530 within ~20 km of the coast due to the inclusion of land-sea breeze forcing,
 531 relative to the case with upwelling winds alone. We posit two explanations
 532 for this perhaps unintuitive result.

533 Firstly, although the Ekman transport ($M_E = \frac{\bar{\tau}_y}{\rho f}$) is the same in both
 534 cases, the cross-shore velocity of the surface Ekman boundary layer ($U_E =$
 535 $\frac{M_E}{H_s}$) is reduced when the land-sea breeze is included, due to diapycnal mixing-
 536 induced deepening of the surface layer (H_s). This leads to a persistently
 537 landward location of the upwelling front when land-sea breeze forcing is ap-
 538 plied, effectively resulting in retention of surface waters and a net warming
 539 in the nearshore (this concept is further discussed in Section 4).

540 The second potential explanation for the nearshore warming due to the
 541 inclusion of diurnal wind variability is that the enhanced diapycnal mixing
 542 leads to a warming of subsurface waters, which are carried by the upwelling
 543 circulation toward the inshore and surface. As it is not immediately apparent
 544 which of these two processes (the location of the upwelling front vs modi-
 545 fied upwelled water) is of leading order in explaining the nearshore surface
 546 warming, Appendix A presents diagnostics designed to reveal the relative
 547 contribution of each. The analysis suggests that the nearshore warming due
 548 to the inclusion of the land-sea breeze is primarily driven by the location of
 549 the upwelling front, while the upwelling of warmer waters due to enhanced
 550 vertical mixing plays a secondary role.

551 3.4. Case study of St Helena Bay

552 We now present the final more comprehensive experiment, being a realis-
 553 tically configured simulation for comparison with the nearshore observations
 554 of Lucas et al. (2014). The model solution is integrated over a 7 day period,
 555 starting on 7 March 2011 from the temperature section shown in Figure 1,
 556 and forced with hourly winds and heat fluxes derived from the 3 km resolu-
 557 tion CSAG WRF model output.

558 Figure 9 compares the observed and modelled temperature across the
 559 mooring array, including summary statistics of model bias, centred root mean
 560 square difference (RMSD) and Pearson's correlation coefficient, as a func-
 561 tion of depth. The model is shown to reproduce the salient features of the

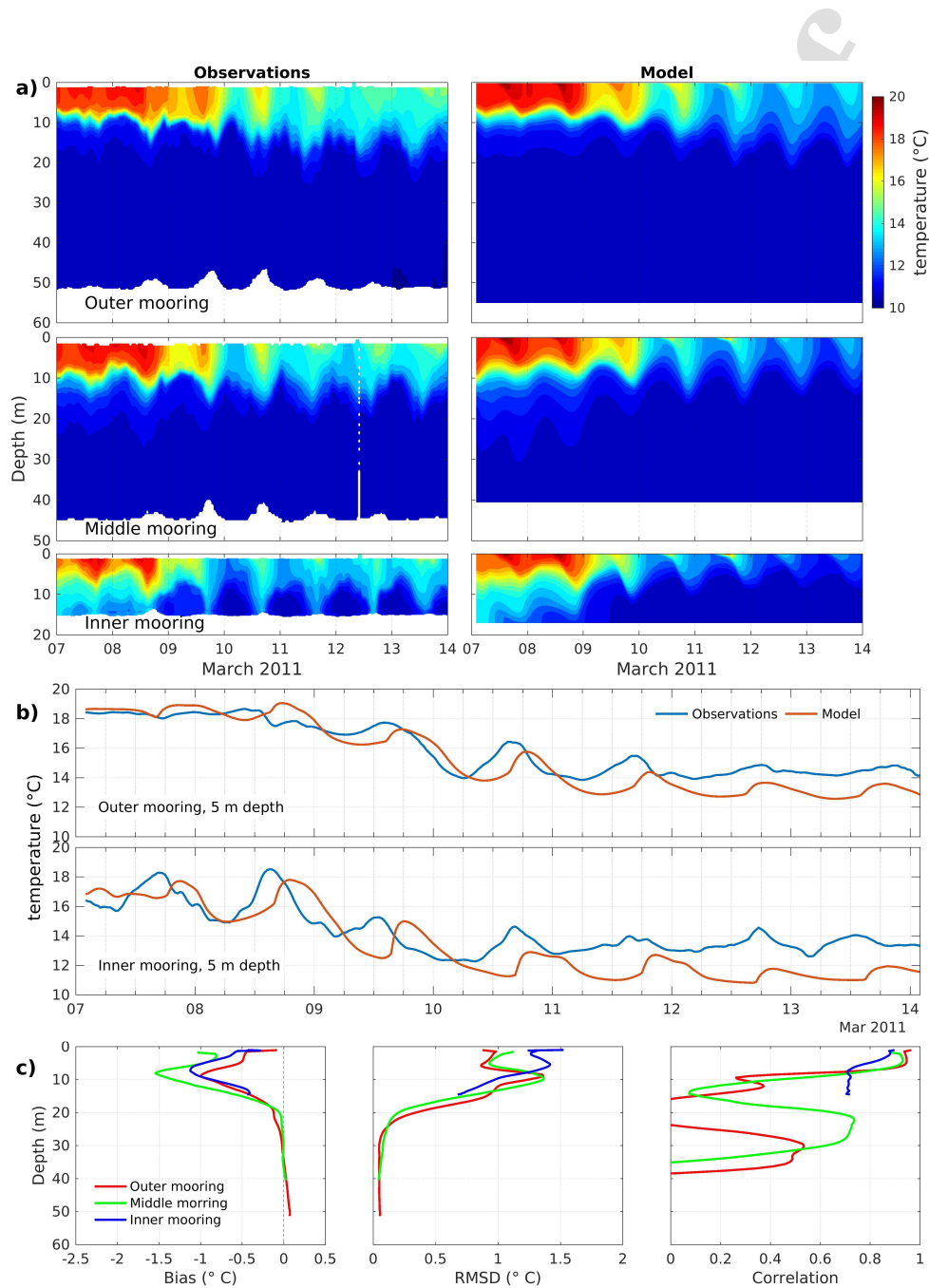


Figure 9: Observed and modelled temperature over an upwelling event accompanied by land-sea breeze forcing from a mooring array in St Helena Bay (Figure 1). (a) Temporal evolution of the observed (left) and modelled (right) temperature profile at the outer (~ 12.7 km offshore), middle (~ 8.1 km offshore) and inner (~ 3.9 km offshore) moorings; (b) time-series of observed and modelled temperature at the outer and inner moorings at 5 m depth; (c) summary statistics for modelled temperature as a function of depth. The model is shown to reproduce the salient features of the temperature observations, characterised by diurnal-inertial variability in surface temperature, large thermocline displacements and a net cooling of surface waters.

562 temperature observations at all three moorings, which are characterised by
 563 significant diurnal-inertial variability in surface temperature and a net cool-
 564 ing of surface waters over the considered period. The analytically configured
 565 2D experiments have highlighted the mechanism whereby near the critical
 566 latitude of 30° S, diurnal-inertial oscillations in the presence of sub-inertial
 567 upwelling produces strong diurnal variability in surface temperature through
 568 repeated advection of strong horizontal temperature gradients over the mea-
 569 surement location. The pumping of the thermocline, characteristic of the
 570 internal wave response as revealed in the analytically configured 2D experi-
 571 ments, is also evident in both the observations and the model. It is expected
 572 that the 2D model would tend to over-estimate thermocline displacements,
 573 as the divergence/ convergence of the cross-shore flow ($\frac{\partial u}{\partial x}$) is compensated by
 574 vertical motion ($\frac{\partial w}{\partial z}$) alone due to the exclusion of alongshore gradients ($\frac{\partial v}{\partial y}$)
 575 in the model. Alongshore velocity gradients however appeared to have played
 576 a minor role over the observation period, as revealed by similar amplitude
 577 (~ 5 m) thermocline displacements in the both model and the observations.

578 The model performance is best at the surface, as reflected by high cor-
 579 relation coefficients at all three moorings (>0.8 for depths shallower than
 580 ~ 5 m, Figure 9c). The significant reduction in temperature correlation near
 581 the base of the thermocline can in part be explained by an over-prediction
 582 in sub-inertial upwelling and the consequent over-prediction in thermocline
 583 lifting, as reflected in a negative near-surface temperature bias across the
 584 mooring array (left panel of Figure 9c). This could be attributed to an over-
 585 prediction in the mean alongshore surface wind stress derived from the CSAG
 586 WRF model, or potentially due to the advection of alongshore variability in
 587 temperature which is explicitly excluded in the model. The low (even nega-
 588 tive) correlations in the subsurface water (>20 m depth) at the middle and
 589 outer moorings are attributed to the lack of a meaningful temperature signal
 590 here over the considered event, as evidenced by the low RMSD in this region
 591 of the water column.

592 The time-series comparisons at 5 m depth (Figure 9b) indicate a phase lag
 593 (~ 6 hrs) in the timing of the diurnal fluctuations in near-surface temperature,
 594 which may reflect shortcomings in the timing of the land-sea breeze in the
 595 CSAG WRF model forcing. One potential explanation for this could be due
 596 to shortcomings in the sea surface temperature (SST) input to the CSAG
 597 WRF model. Upwelling is known to impact the timing of the land-sea breeze
 598 (Seroka et al., 2018), while the WRF model uses daily SST fields derived from
 599 satellite observations which contain systematic errors in upwelling regions

600 (Meneghesso et al., 2020) and do not capture super-diurnal variability in
601 SST's.

602 The observed and modelled cross-shore component of velocity for the
603 outer and inner moorings are compared in Figure 10 (ADCP data from the
604 middle mooring were mostly missing over this period and is therefore not
605 shown). While the salient features of the vertical structure of the currents
606 are reproduced in the model, the time-series comparisons (Figure 10b) again
607 reveal phase differences in the diurnal-inertial oscillations, where the model
608 lags the observations by ~ 6 hrs at both the outer and inner moorings.

609 Despite the presence of the time lag, correlation coefficients at the outer
610 mooring are between ~ 0.4 and ~ 0.6 for the subsurface currents, while a
611 maximum of ~ 0.7 is attained for the near-surface currents (Figure 10c). The
612 model correlation coefficients drop to negative values in the region of the
613 thermocline, where the 180° phase shift between the surface and subsurface
614 currents is observed. Poor correlation in this region is not surprising, as any
615 misrepresentation in the depth of the 180° phase shift will have a notable
616 impact on the correlations in this region of the water column. Current cor-
617 relations are lower at the inner mooring (between ~ 0.2 and ~ 0.4), where the
618 amplitude of the diurnal-inertial signal is weaker. The modelled currents at
619 this mooring appear to be hampered by the signature of an over-prediction in
620 sub-inertial upwelling circulation, as evidenced by negative (positive) cross-
621 shore current biases in the surface (subsurface) (Figure 10c). These biases
622 are in agreement with the cool surface temperature bias visible in Figure 9.

623 4. Discussion and conclusions

624 In this paper we have employed novel 2D model experiments to gain
625 insight into the cross-shore processes governing the response of coastal up-
626 welling systems to land-sea breeze forcing. In addition to testing model
627 sensitivity to relevant variables in analytically configured experiments, a re-
628 alistically configured experiment has been compared with three nearshore
629 moorings orientated perpendicular to the local coastline in St Helena Bay,
630 located in the Southern Benguela Upwelling System (Lucas et al., 2014).
631 Our analytically configured experiments have focussed on the response of
632 upwelling systems at the critical latitude of 30° N/S, where the diurnal pe-
633 riodicity of the land-sea breeze is resonant with the inertial frequency and
634 the locally forced response is maximised (Craig, 1989; Simpson et al., 2002;
635 Fearon et al., 2020). Given the mooring array latitude of 32.3° S, the observa-

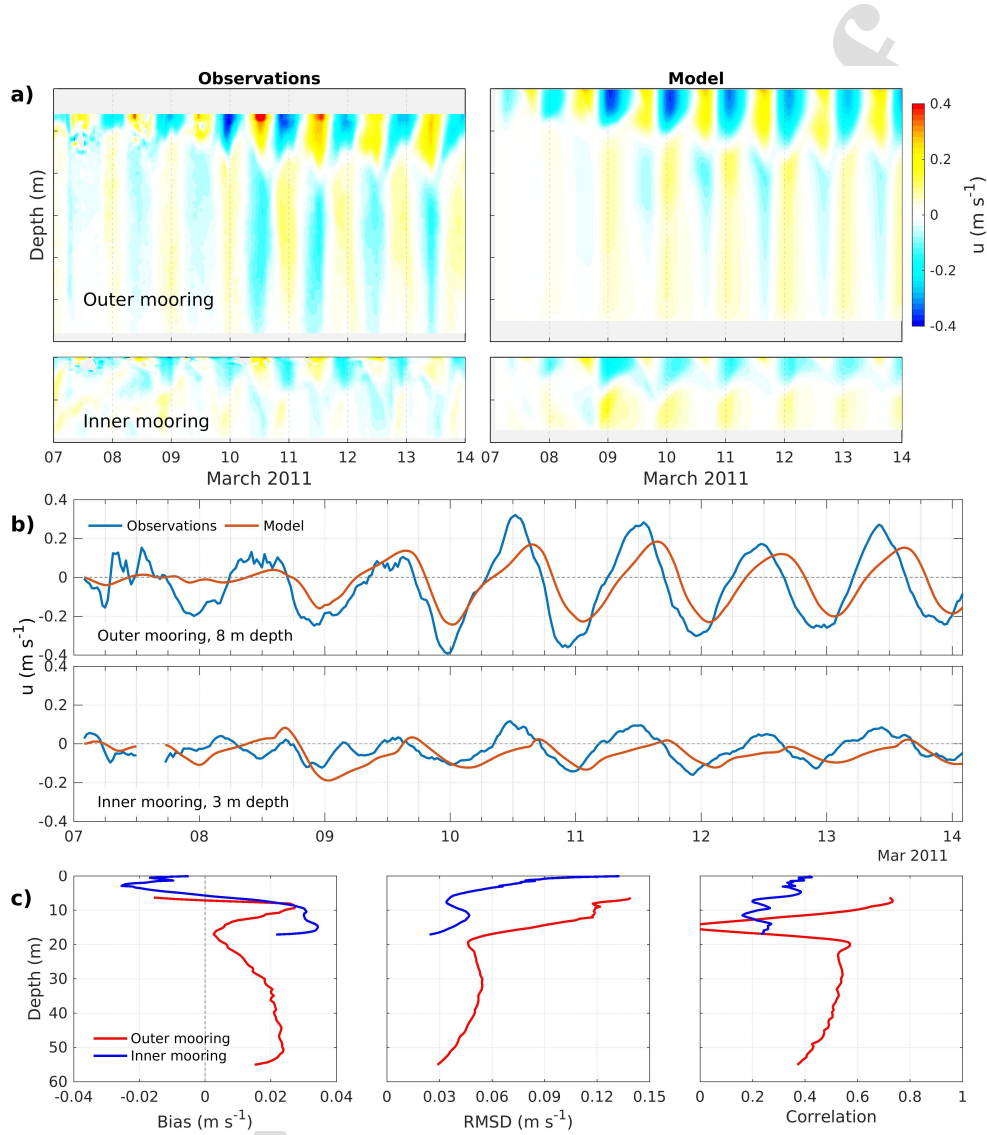


Figure 10: Observed and modelled cross-shore velocity (u) over an upwelling event accompanied by land-sea breeze forcing from a mooring array in St Helena Bay (Figure 1). (a) Temporal evolution of the observed (left) and modelled (right) cross-shore velocity profile at the outer (~ 12.7 km offshore) and inner (~ 3.9 km offshore) moorings; (b) time-series of observed and modelled cross-shore velocity at the outer and inner moorings at 8 m and 3 m depths, respectively; (c) summary statistics for modelled cross-shore velocity as a function of depth. ADCP data from the middle mooring were mostly missing over this period and is therefore not shown. The model is shown to reproduce the salient features of the current observations, characterised by diurnal-inertial oscillations in surface and subsurface layers with a 180° phase shift between the two.

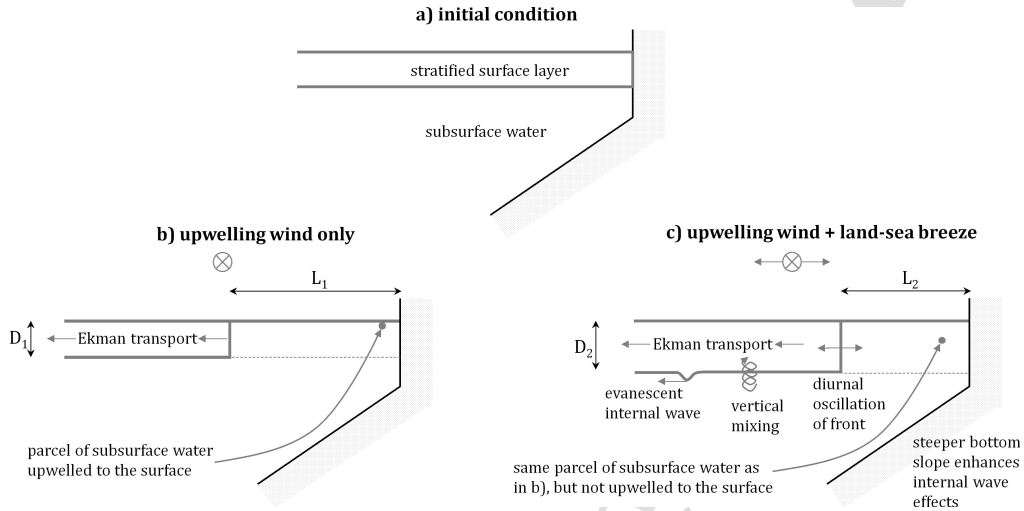


Figure 11: Summary of key processes resulting from land-sea breeze forcing in upwelling systems near the critical latitude of 30° N/S.

636 tions provide a particularly relevant ground truth for the presented 2D model
 637 configuration. Notwithstanding model shortcomings which include an over-
 638 prediction in sub-inertial upwelling and notable phase lags in the modelled
 639 diurnal-inertial response, the favourable comparison with the observations
 640 provides some confidence in the dynamics elucidated by the analytically con-
 641 figured experiments.

642 The presented results are relevant for all regions under the influence of
 643 land-sea breeze forcing near the critical latitude, which includes all four of the
 644 major Eastern Boundary Upwelling Systems (EBUS). The relevance of the
 645 results naturally diminishes away from the critical latitude, where the locally
 646 forced resonance phenomenon becomes weaker (e.g. Figure 5 and Figure 10
 647 of Fearon et al., 2020). Hyder et al. (2011) argue the region of influence to
 648 lie between latitudes of 23° and 40° N/S. The key processes elucidated by
 649 the experiments presented in this paper are summarised in Figure 11 and
 650 discussed below.

651 4.1. Separation of the forced and internal wave responses

652 Contrasting results from the 2D model with those of the 1D model pre-
 653 sented in Fearon et al. (2020) has allowed us, for the first time, to identify
 654 the respective roles of the locally forced response (as approximated by the
 655 1D model) and the internal wave response to land-sea breeze forcing near the

656 critical latitude. While the focus of this paper is on coastal upwelling sys-
657 tems, the findings of these purely land-sea breeze forced experiments would
658 apply to any two layer coastal systems near the critical latitude.

659 As the locally forced response to the land-sea breeze is maximised at
660 the critical latitude of 30° N/S, here we also find the maximum amplitude
661 thermocline displacements due to convergence/ divergence of the forced oscil-
662 lations at the land boundary. Although the internal waves are evanescent at
663 the critical latitude, in agreement with theory and the numerical experiments
664 of Zhang et al. (2010), they provide an additional source of shear-driven di-
665 apycnal mixing. In the considered configuration, the effect is maximised at
666 a distance of ~ 25 km from the wave generation zone at the land boundary,
667 where diapycnal mixing can be $\sim 20\%$ greater than that due to the forced
668 response alone. The influence of the internal waves extends to ~ 50 km from
669 the land boundary, offshore of which the 1D and 2D model solutions are in
670 near-perfect agreement.

671 Sensitivity tests have revealed the importance of the bottom slope in
672 governing the processes driving the cross-shore variability in near-inertial
673 motions. The steepness of the bottom slope determines the rate at which
674 the locally forced oscillations are dampened toward the coast, thereby influ-
675 encing the convergence/ divergence of the surface currents and thermocline
676 pumping. Internal wave generation at the coast is therefore dampened by a
677 gradually sloping bottom. This is in agreement with previous 2D primitive
678 equation experiments over the gradually sloping Texas-Louisiana shelf, where
679 non-linear advection terms were of secondary importance and the first order
680 cross-shore surface elevation gradient was found to play a dominant role in
681 the cross-shore variability in near-inertial oscillations (Chen and Xie, 1997).
682 In contrast, a simple two-layer coastal wall model has been used to identify
683 the internal wave generation at the boundary as the driving mechanism for
684 explaining the drop-off in near-inertial energy toward the coast (Shearman,
685 2005). Our sensitivity experiments serve to reveal how the bottom slope
686 governs which of these processes (first order surface elevation gradient vs
687 internal wave generation) dominate the reduction in near-inertial motions
688 toward the coast. Steeper bottom slopes are required to enhance the internal
689 wave response and allow for elevated diapycnal mixing over and above what
690 would be expected from the forced response alone.

691 *4.2. The influence of the land-sea breeze on sub-inertial upwelling*

692 The inclusion of a sub-inertial alongshore wind stress in the 2D model
 693 experiments has allowed for further reflection on the role of diurnal-inertial
 694 resonance within the context of sub-inertial upwelling dynamics. Our re-
 695 sults reveal how the combined effect of sub-inertial upwelling and inertial
 696 oscillations can produce pronounced diurnal variability in nearshore surface
 697 temperatures, driven by horizontal diurnal advection of strong cross-shore
 698 temperature gradients, which are ever-present in upwelling systems. Local
 699 land-sea breeze forcing has been identified as playing a key role in diurnal
 700 temperature variability in a number of in-situ observations in EBUS (Ka-
 701 plan et al., 2003; Woodson et al., 2007; Bonicelli et al., 2014; Walter et al.,
 702 2017) and is evident in the nearshore observations of Lucas et al. (2014)
 703 (Figure 9). The 2D model experiments presented here clearly demonstrate a
 704 driving mechanism of this phenomenon, although it is noted that solar irra-
 705 diance is an additional mechanism which would further contribute to diurnal
 706 variability in sea surface temperatures. While not explicitly considered in
 707 this paper, the impact of solar irradiance on surface temperatures over the
 708 presented observation period can be inferred from the 1D model results at
 709 the outer mooring (Figure 6 of Fearon et al., 2020), which included realistic
 710 surface heat fluxes. Comparison of the modelled temperature at the outer
 711 mooring from the 2D model (Figure 9a) with that of the 1D model suggests
 712 that the advection of horizontal gradients was the dominant mechanism in
 713 driving the diurnal sea surface temperature variability over the considered
 714 observation period.

715 Our simulations have further highlighted how deepening of the thermo-
 716 cline due to diurnal-inertial resonance near the critical latitude can lead to a
 717 reduction in sub-inertial offshore advection of the surface Ekman boundary
 718 layer, as illustrated in Figure 11. As offshore Ekman transport is the same
 719 in both Figure 11b and Figure 11c, volume conservation dictates that D_1L_1
 720 $= D_2L_2$, and therefore the landward offset of the upwelling front will scale
 721 linearly with the increased depth in the surface Ekman boundary layer. For
 722 example, if a mean alongshore wind drives a 15 m deep Ekman boundary
 723 layer 10 km offshore, the same wind will drive a 20 m deep Ekman bound-
 724 ary layer 7.5 km offshore ($15 \div 20 \times 10$). Shallower surface mixed layers elicit
 725 higher amplitude oscillations in the surface layer and greater mixing (Fearon
 726 et al., 2020), implying that regions prone to the development of shallow
 727 surface layers, such as retention zones in the lee of capes (e.g. Graham and
 728 Largier, 1997; Oliveira et al., 2009), are likely to be most affected by this pro-

729 cess. Advective losses during active upwelling can contribute to a reduction
730 in productivity within upwelling systems, while the retention of upwelled
731 waters is important for the accumulation of high biomass coastal blooms
732 (Pitcher et al., 2010). Our results imply that the land-sea breeze may play
733 a contributing role in surface water retention in these areas through reduced
734 offshore advection of the surface layer during upwelling events.

735 The retention of surface waters due to the inclusion of land-sea breeze
736 forcing has been shown to drive a net warming of nearshore surface waters,
737 primarily due to the landward displacement of the upwelling front, while
738 modification of subsurface waters through enhanced diapycnal mixing plays
739 a secondary role. Offshore of the influence of the upwelling front, surface
740 waters are cooled through the entrainment of subsurface waters via diapyc-
741 nal mixing. These results suggest that the misrepresentation of the land-sea
742 breeze in global and regional models of EBUS may contribute to nearshore
743 temperature biases in these models. Such biases are largely attributed to er-
744 rors in the representation of the low frequency alongshore wind stress in the
745 atmospheric forcing products (e.g. Richter, 2015; Small et al., 2015), while
746 the influence of the land-sea breeze is over-looked as a potential source of
747 systematic error. Our results suggest that the improvement in the spatial
748 and temporal representation of the land-sea breeze could be important in al-
749 levating this source of systematic bias, particularly near the critical latitude
750 and in areas susceptible to the development of shallow surface mixed layers.

751 We note that our results have been limited to short duration (7 day)
752 experiments with a simplified 2D model. While these experiments have been
753 useful to elucidate event-scale dynamics, further work is required to assess
754 the overall role of these processes within EBUS, where Ekman dynamics
755 are also counteracted by eddy effects (Thomsen et al., 2021). Inter-annual
756 simulations of realistically configured 3D models of EBUS, which both include
757 and exclude land-sea breeze forcing, are expected to provide useful insights to
758 this end. Such regional simulations are typically forced at the surface by an
759 atmospheric model which is itself forced by sea surface temperatures which do
760 not capture land-sea breeze effects on the upwelling front. This shortcoming
761 was invoked as a possible source of the phase lag error in the realistically
762 configured simulation (Section 3.4). Previous studies have identified coastal
763 upwelling as influencing the land-sea breeze both through an earlier onset and
764 enhanced intensity (Clancy et al., 1979; Seroka et al., 2018), which may in
765 turn have a positive feedback on upwelling favourable winds (Franchito et al.,
766 1998). This paper has further explored how the land-sea breeze may impact

767 the location of the upwelling front near the critical latitude; processes which
768 themselves appear likely to influence the land-sea breeze. It is suggested that
769 two-way coupled ocean-atmosphere experiments may prove to be particularly
770 insightful in exploring this two-way feedback near the critical latitude.

771 *4.3. Improved understanding of the St Helena Bay observations*

772 The results of this paper provide an opportunity for further reflection on
773 the processes governing the nearshore observations of Lucas et al. (2014).
774 While the forced response, as approximated by the 1D model, reproduces
775 many of the salient features of the observations at the offshore mooring (Fig-
776 ure 6 of Fearon et al., 2020), the 1D model was shown to over-predict the
777 deepening of the thermocline through diapycnal mixing, under-predict net
778 cooling of surface waters and lack sufficient diurnal-inertial variability in both
779 surface temperature and vertical displacement of the thermocline. The real-
780 istically configured 2D model has significantly alleviated these shortcomings
781 (Figures 9 and 10), suggesting that the analytically configured 2D experi-
782 ments may be useful in interpreting the observations. Figure 6 confirms that
783 for a 1:200 bottom slope (the approximate slope across the mooring array)
784 and an offshore distance of just ~ 12.7 km, an over-prediction in the ampli-
785 tude of the surface oscillation and consequent deepening of the thermocline
786 is to be expected in the 1D model. Here, the 2D model experiments reveal
787 the ~ 5 m amplitude thermocline displacements to be the signature of in-
788 ternal waves generated at the land boundary. The inclusion of sub-inertial
789 upwelling circulation in the 2D model alleviates the 1D model shortcomings
790 of an under-prediction of net cooling of surface waters (although upwelling is
791 over-predicted in the 2D model), and provides the mechanism for generating
792 the observed diurnal variability in surface temperatures, as already discussed.

793 Lucas et al. (2014) identified vertical mixing-driven nutrient flux to be
794 of high importance in governing observed phytoplankton growth across the
795 mooring array. While the 2D model confirms shear-induced diapycnal mix-
796 ing to be a feature over the mooring array, an important consequence of
797 the presented experiments is that the region of maximum diapycnal mixing
798 is in fact expected to be located offshore of the observations. It therefore
799 seems plausible that the effect of the land-sea breeze on the phytoplankton
800 response may well have been further enhanced offshore of the the mooring
801 array. This suggests that our understanding of event-scale phytoplankton
802 dynamics in the region would greatly benefit from high-frequency observa-
803 tions of the water column which extend across the full width of St Helena

804 Bay.

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820 Appendix A. Diagnostics explaining warming of nearshore surface 821 waters in response to land-sea breeze forcing

822 Figure 8 indicates nearshore surface warming due to the inclusion of land-
 823 sea breeze forcing, while two processes are identified as possible drivers of this
 824 result:

- 825 1. the landward location of the upwelling front due to a deepened Ekman
 826 boundary layer and
- 827 2. the warming of subsurface water through diapycnal mixing, which is
 828 then upwelled to the surface.

829 In this appendix we present diagnostics designed to reveal which of these
 830 two processes is of leading order in explaining the nearshore surface warming.

831 The contribution of the location of the upwelling front (blue line in Fig-
 832 ure A.12b) is computed as follows: the cross-shore temperature profile from
 833 the surface layer of the simulation forced by an upwelling wind alone is ex-
 834 tracted at each 30 min time-step; a landward offset to this profile is then
 835 applied according to the difference in the upwelling front locations shown in
 836 Figure A.12a; the difference between the original and the offset cross-shore

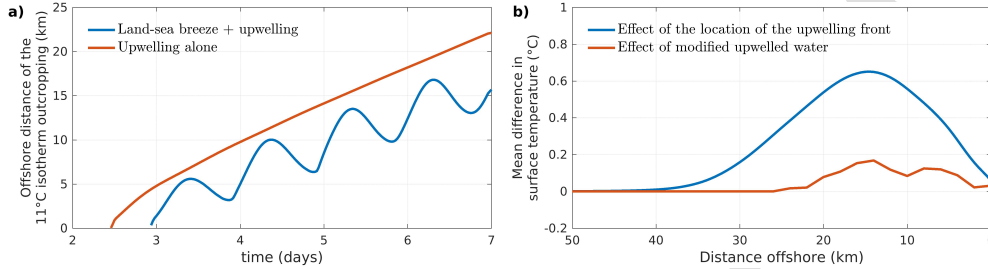


Figure A.12: Diagnostics used to explain the warming of nearshore surface waters in response to the inclusion of the land-sea breeze, as shown in Figure 8. (a) Time-series of the offshore distance of the 11° C isotherm in the surface layer of the model (used as a proxy for the upwelling front) for the simulations shown in Figure 8. (b) Estimates of the relative contribution of the two processes (the location of the upwelling front vs modified upwelled water) which explain the warming of nearshore waters due to the inclusion of the land-sea breeze.

837 temperature profiles is then computed. The effect shown in Figure A.12b is
 838 the mean temperature difference over the entire 7 day simulation.

839 The contribution of the upwelling of warmer subsurface waters is com-
 840 puted through the use of CROCO's online Lagrangian floats module. Both
 841 simulations (i.e. excluding and including land-sea breeze forcing) are ini-
 842 tialised with neutrally buoyant Lagrangian floats from the depth of the 11° C
 843 isotherm to the bottom in vertical increments of 1 m and horizontal incre-
 844 ments of 500 m (the horizontal grid resolution), thereby covering subsurface
 845 waters over the entire model domain. The temperature at the location of
 846 the floats, as interpolated from the Eulerian model, is saved at 30 min in-
 847 tervals over each simulation. Temperature differences between the Eulerian
 848 model at the location of the Lagrangian floats and the initial temperature of
 849 Lagrangian floats are used to quantify how subsurface water is modified by
 850 land-sea breeze-driven diapycnal mixing over the upwelling simulations. The
 851 red line in Figure A.12b) is computed as follows: all Lagrangian floats in the
 852 upper 5 m of the water column are identified at each 30 min time-step and
 853 taken to represent upwelled water; the difference between the temperature of
 854 the Eulerian field at the location of the floats and the original temperature
 855 of the floats is computed (quantifying how the upwelled water has been mod-
 856 ified through diapycnal mixing); the offshore distance of the floats are used
 857 to construct a cross-shore profile of the surface temperature difference, using
 858 2 m bins in the cross-shore direction and averaging the temperature differ-
 859 ence from all 'upwelled' Lagrangian floats contained within each 2 m bin.

860 The effect shown in Figure A.12b is the mean temperature difference over
 861 the entire 7 day simulation forced by a combination of land-sea breeze and
 862 upwelling winds, over and above that computed from the simulation forced
 863 by an upwelling wind alone. This approach is designed to capture how the
 864 temperature of subsurface water has been modified by enhanced diapycnal
 865 mixing offshore of the upwelling front, prior to upwelling.

866 Figure A.12b suggests that the nearshore warming due to the inclusion of
 867 the land-sea breeze is primarily driven by the location of the upwelling front,
 868 while the upwelling of warmer waters due to enhanced vertical mixing plays
 869 a secondary role.

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- Latitudinally dependent diurnal wind-driven internal waves enhance vertical mixing
- Vertical mixing drives surface layer retention and coastal warming during upwelling
- Diurnal wind-driven oscillations of the upwelling front explain observed temperatures
- Bottom slope steepness controls internal wave generation at the coast

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Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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