

**Title:** Pushing and pulling an algal bloom: physical controls of diel variability in nearshore phytoplankton communities

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# 1 Highlights

## 2 **Pushing and pulling an algal bloom: Physical controls of diel vari-** 3 **ability in nearshore phytoplankton biomass**

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- 5     • Nearshore phytoplankton assemblages can undergo fast short-term vari-  
6       ations not exclusively imputable to population growth.
- 7     • These diel changes in microalgal cell abundance are the source of en-  
8       hanced water discoloration and other harmful effects.
- 9     • Wind and buoyancy-driven flows were identified as key drivers of the  
10       observed fluctuations in Palma Beach (NW Mediterranean).
- 11    • A 1D advection-diffusion model successfully diagnoses the mechanisms  
12       underlying the daily dynamics of the proliferation of coastal microalgae.

13 Pushing and pulling an algal bloom: Physical controls  
14 of diel variability in nearshore phytoplankton biomass

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16 **Abstract**

High-biomass microalgal blooms frequently occur in littoral environments worldwide, often causing noxious effects on aquatic ecosystems and coastal communities. Here, we combine field observations and a simple retention-dispersion model to disentangle the short-term ( $\sim$  hours) environmental drivers shaping the nearshore dynamics of such outbreaks. Temperature, salinity, fluorescence, current velocities, and meteorological variables were measured in the nearshore waters of a coastal location in Mallorca (Balearic Islands) during the summer of 2018. Daily averages from field data were used to adjust wind and buoyancy flow variations into a one-dimensional advection-diffusion model in the cross-shore direction. Results reveal that the interplay between wind forcing and cross-shore density gradients drives an alternating retention dispersion mechanism, effectively explaining the observed diel chlorophyll variability within the nearshore boundary. This simplified model captures the primary dynamics of the bloom, isolating key factors that

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influence its behavior and offering practical insights for coastal water quality monitoring and management.

<sup>17</sup> *Keywords:* Phytoplankton, Chlorophyll, Wind stress, Thermo-haline flow,  
<sup>18</sup> Advection - diffusion model, Ground water, Littoral dynamics

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## 19 1. Introduction

20 Phytoplankton are key unicellular organisms that thrive in fresh and ocean  
21 waters, playing a vital role in the production of organic matter and oxygen,  
22 while also contributing to the regulation of global CO<sub>2</sub> levels and the Earth's  
23 climate. However, under favorable environmental conditions, phytoplank-  
24 ton can multiply and accumulate, causing negative effects on the ecological  
25 balance of aquatic ecosystems. These high biomass Harmful Algal Blooms  
26 (HABs) cause significant socio-economic impacts on local communities and,  
27 in some cases, toxicity effects may arise [1] [2].

28 Phytoplankton and, particularly some species of microalgae, often find suit-  
29 able environments to proliferate in coastal waters, where enclosed areas like  
30 estuaries and bays offer suitable conditions for calm waters with high nutri-  
31 ent availability from terrestrial inputs [3]. In more exposed locations, coastal  
32 HABs may occur as massive cell accumulations extending along a nearshore  
33 stripe producing a perceivable cross-shore gradient in water discoloration  
34 [4].

35 Phytoplankton growth in the coastal boundary is favored by terrestrial nu-  
36 trient sources delivered to nearshore waters by rivers, urban outlets, or sub-  
37 marine groundwater discharges (SGD), which introduce an abundance of nu-  
38 trients that are otherwise scarce in seawater [5] [6]. The prime force driving  
39 circulation in the coastal boundary zone is the wind; however, the effects of  
40 stratification produced by terrestrial seeps and warming during the summer

41 period can become very important [7] [8]. The diurnal sea breeze, which  
42 on many coasts is seasonal and locally predictable in its occurrence, rep-  
43 resents a recurrent force affecting the cross-shore distribution of material  
44 and organisms suspended in the water column [9]. Also, fresh and brack-  
45 ish water discharges in the nearshore and diurnal warming generate vertical  
46 stratification and convective horizontal exchange flows driven by destabiliz-  
47 ing buoyancy fluxes that play an important role in the transport of nutri-  
48 ents, pollutants, and chemical substances between the littoral and pelagic  
49 regions also modulating nearshore plankton communities [10]. The convec-  
50 tively driven horizontal flows are the result of a density difference in the  
51 horizontal direction. Solar radiation during the day leads to warmer water  
52 in the shallows than in the adjacent deeper regions, and this developing con-  
53 trast in temperature between shallow and deep waters produces variations  
54 in water density that generate convective water exchange [11] [12]. Likewise,  
55 SGD seeps along the shoreline produce density contrast due to freshwater  
56 influx. These density-driven buoyant plumes modify nearshore circulation  
57 patterns influencing phytoplankton distribution [13].

58 Despite their important environmental and social consequences, the mech-  
59 anisms driving phytoplankton variability in nearshore waters and the even-  
60 tual occurrence and evolution of HABs are difficult to fully understand due  
61 to their short timescales of variation (hours) and intrinsic multifactorial na-  
62 ture in which physical, biological, and geochemical processes intervene. High

63 biomass outbreaks may occur that are too rapid and intense to be explained  
64 by reported microalgal cell growth rates [14], suggesting that physical ac-  
65 cumulation mechanisms may be major drivers of these events. For exam-  
66 ple, coastal blooms in the Mediterranean Sea are typically dominated by di-  
67 noflagellates showing individual growth rates in the range of 0.4 - 0.8 day<sup>-1</sup>  
68 [16] [15]. However, it is not uncommon for daily cell abundance to vary by  
69 more than threefold [17].

70 Circulation in the nearshore can either favor or hamper high-biomass HAB  
71 development by either accumulating or dispersing algal populations. For ex-  
72 ample, wind-driven circulation has been observed to maintain and generate  
73 algal blooms both in the sea and in lakes by downwind transport and ac-  
74 cumulation of HAB-producing cells [14] [18] [19]. Likewise, the presence  
75 of intense temperature and salinity gradients at sea fronts can represent a  
76 boundary favoring phytoplankton accumulation [20] [21] [22]. In the coastal  
77 boundary zone, these accumulation mechanisms operate at short timescales  
78 (hours). However, their influence on the dynamics of nearshore phytoplank-  
79 ton remains to be fully explored While most field studies on HAB dynamics  
80 have focused on mesoscale (kilometers to hundreds of kilometers) or regional  
81 processes, much less attention has been paid to fine-scale processes occur-  
82 ring at sub-kilometer spatial resolution and over short timescales (hours).  
83 However, these small-scale processes can be critical for understanding the  
84 mechanisms underlying rapid biomass accumulation in coastal areas.

85 Here, we examine the dynamic interplay between the physical forcings that  
86 drive the cross-shore displacement of the nearshore water strip and its in-  
87 fluence on phytoplankton accumulation at Palma Beach (Balearic Islands)  
88 during a small algal biomass outbreak. We focus on the understanding of  
89 the diel variability patterns of phytoplankton as determined by wind-forced  
90 transport and buoyancy currents. Data from three moorings and repetitive  
91 sampling in nearshore waters and a simple advection-diffusion 1D model are  
92 used to interpret the retention-dispersion mechanisms regulating the short-  
93 term (hours) phytoplankton variability.

## 94 **2. Methods**

### 95 *2.1. Site description*

96 The case study site, Palma Beach ( $39^{\circ}31'11.5''$  N,  $2^{\circ}44'06.3''$  E), is located  
97 on the inner coast of Palma Bay, southern Mallorca In summer, it features  
98 (Balearic Islands), and is characterized by a semi-enclosed morphology with  
99 high anthropogenic influence. During the summer, the area exhibits calm  
100 hydrodynamic conditions driven by weak sea breeze winds, negligible tidal  
101 forcing, minimal wave activity, and a gently sloping bathymetry ( $\sim 10$  m per  
102 km) (  $10$  m per km), which lacks breaker zones and promotes water mass  
103 retention (Jordi et al., 2011). This 5 km-long beach is subject to intense hu-  
104 man pressure from urban development and seasonal tourism, both of which  
105 contribute to environmental stress. A notable feature affecting coastal water  
106 quality is the seepage of a nutrient-enriched groundwater from the coastal

107 aquifer often fuels dinoflagellate blooms [24] along the beach, which intro-  
 108 duces elevated levels of dissolved nitrogen and trace elements such as iron  
 109 into the littoral zone [6][24]. These nutrient inputs create fertile nearshore  
 110 conditions that, combined with elevated summer sea temperatures, enhance  
 111 phytoplankton productivity and increase the likelihood of microalgal bloom  
 112 formation. Such blooms are typically short-lived yet recurrent, occurring  
 113 from late May through September, and align with seasonal bloom dynamics  
 114 observed at other Mediterranean coastal sites [17][43]

## 115 2.2. Field data

116 Temperature, salinity, current velocity, and chlorophyll ( $\theta$ ,  $S$ ,  $u$ , and  $Chl$ )  
 117 were monitored over one week (Julyb 19-25, 2018). Cross-shore transects of  
 118 near-surface (0.25 m) fluorescence, salinity, and temperature were measured  
 119 every 3 hours from the shoreline to 600 m offshore using a flow-through sys-  
 120 tem on a small boat [4] (Fig. 1). Seawater samples were collected at three  
 121 points (P1, P6, P9; Fig. 1) using 0.6 L Niskin bottles, and chlorophyll con-  
 122 centrations were determined by filtering, allowing fluorescence calibration  
 123 (see Appendix A.1). Wind ( $W$ ) data were obtained from an oceanographic  
 124 buoy 3.5 km offshore (SOCIB). Cross-shore currents were measured with  
 125 bottom-mounted ADCPs (Nortek HR Aquadopp) attached to poles fixed on  
 126 the seabed at three locations (M1, M2, M3; Fig. 1), obtaining current profiles  
 127 from  $\sim 0.15$  m below the surface to the seafloor every 3 seconds [25].

### 128 2.3. A minimal model for the fast nearshore bloom dynamics

129 To disentangle the mechanisms behind the *Chl* patterns observed during the  
 130 field survey, we developed a simple toy model to capture the key environ-  
 131 mental processes driving the fast dynamics (on the scale of hours) in the  
 132 nearshore zone, aiming at minimal parametrization [26]. The phytoplankton  
 133 standing stock (*Chl*) is expected to evolve over time due to multiple processes  
 134 [27]:

$$135 \quad \frac{\partial Chl}{\partial t} = \text{growth} - \text{grazing} + \text{diffusion}_{x,y,z} + \text{advection}_{x,y,z} - \text{sinking}_z \quad (1)$$

136 where each term acts over different timescales. Reported *in situ* growth  
 137 rates of dinoflagellates are generally  $< 1.0 \text{ day}^{-1}$  [16], with grazing pressure  
 138 typically comparable to or lower than this value [28]. To indirectly constrain  
 139 phytoplankton growth rates during the field survey, we refer to NCP-based  
 140 estimates from Basterretxea (2024), which reported rates of  $0.16\text{-}0.25 \text{ day}^{-1}$ ,  
 141 indicative of a community not undergoing rapid proliferation. We thus expect  
 142 growth and grazing to operate on timescales  $> 1 \text{ day}$  (doubling or halving  
 143 times). Since the most significant biomass fluctuations in our data occur on  
 144 the timescale of hours (Fig. 2a), with minimal variance between days and  
 145 a roughly constant total biomass standing stock throughout the sampling  
 146 period ( -4% daily biomass loss, Fig. 2a, inset), growth and grazing are ex-  
 147 cluded from the toy model. CTD data reveal vertically homogeneous profiles  
 148 of  $\theta$ ,  $S$ , and *Chl* (see Appendix Appendix A.2), indicating that wave-driven

149 turbulence in these shallow waters maintains well-mixed conditions, thereby  
 150 limiting vertical processes such as sinking or behavioral adaptations (e.g.,  
 151 diel vertical migration). These simplifications are supported by previous  
 152 Empirical Orthogonal Function (EOF) analysis conducted on the dataset  
 153 [25], which found that over 90% of the daily *Chl* variance was driven by  
 154 variations in environmental factors, particularly cross-shore currents, cross-  
 155 shore wind, and temperature. Alongshore variations, although necessary to  
 156 solve the hydrodynamics correctly, are thus also excluded from the toy model.

157  
 158 Overall, we hypothesize that the rapid daily fluctuations ( $\sim$  hours) in nearshore  
 159 biomass are primarily driven by horizontal processes resulting from a dynam-  
 160 ical equilibrium between advection and diffusion in the cross-shore direction,  
 161 with growth and grazing modulating the system on longer timescales ( $\geq 1$   
 162 day). The toy model is finally based on three assumptions:

- 163 1. All physical drivers producing advective flows follow a daily periodicity
- 164 2. The nearshore flow can be reduced to a 1D water mass being pulled  
 165 back and forward from/to the shoreline, and still capture the main  
 166 accumulation-relaxation dynamics
- 167 3. Phytoplankton biomass (determined as *Chl*) behaves as a well-mixed  
 168 passive tracer, excluding relevant cell growth and/or grazing, or be-  
 169 havioural aspects (e.g. dinoflagellates diel vertical migration).

170 The model's first assumption is supported by field data, which show a clear

171 daily periodicity in wind  $W$  (Fig 2a, blue line), temperature  $\theta$  and salinity  $S$   
 172 (Fig 2c-d). Moored instruments confirm a similar periodicity in the current  
 173 data (Fig. 2b for M1; Fig. A.9 for M2 and M3) and reveal a vertically  
 174 homogeneous flow. The absence of vertical structures supports the hypothesis  
 175 of a well-mixed system, as expected in shallow waters where wave-breaking-  
 176 induced turbulence promotes vertical mixing, with the primary dynamics  
 177 occurring in the cross-shore direction.

#### 178 2.4. Building the toy model

179 The  $Chl$  retention-release dynamic along the cross-shore direction is modelled  
 180 with a one-dimensional advection-diffusion equation:

$$181 \quad \frac{\partial Chl(t, x)}{\partial t} = D \frac{\partial^2 Chl(t, x)}{\partial x^2} + u(t, x) \frac{\partial Chl(t, x)}{\partial x} \quad (2)$$

182 where  $x$  is the longitudinal direction along the transect,  $t$  is time,  $u(t, x)$  are  
 183 the cross-shore currents, and  $D$  is an effective diffusion coefficient accounting  
 184 for the mixing processes in the surf zone [30, 31]. The advective term is  
 185 expressed as a linear combination of the main environmental forcings acting  
 186 on the system (wind, salinity and temperature gradients):

$$187 \quad u(x, t) = u_W(x, t) + u_S(x, t) + u_\theta(x, t) \quad (3)$$

188 where  $u_W$  is the wind-driven flow,  $u_S$  is the current component due to the  
 189 fresher water input along the shoreline, and  $u_\theta$  is the current driven by cross-

190 shore temperature gradients. The advective term omits tidal currents, as  
 191 they are expected to play a negligible role in the system's dynamics (see  
 192 [Appendix A.4](#) for details).

193 As shown in Fig [2a](#), wind at the study site followed a daily cycle, peaking at  
 194 5 m/s onshore in the afternoon and dropping and inverting to 2 m/s offshore  
 195 at night. Assuming a linear relationship between the wind speed  $W$  and  
 196 the advected current  $u_W = 5 \cdot 10^{-3} \cdot W$  [\[19\]](#), we expected wind-driven flows  
 197 ranging from 2.5 cm/s onshore to 1 cm/s offshore.

198 The horizontal gradients in  $\theta$  and  $S$  create a buoyant flow driven by density  
 199 differences, where the time required for the flow to develop and its magnitude  
 200 depend on the steepness of the density gradient. Following [\[32\]](#), we assume  
 201 that the horizontal current developing in response to a horizontal density  
 202 gradient  $\partial\rho/\partial x$  is given by the unsteady inertia of water:

$$203 \quad \rho_0 \frac{\partial^2 u}{\partial t \partial z} = g \cdot \frac{\partial \rho}{\partial x} \quad (4)$$

204 where  $\rho$  is the water density,  $\rho_0$  its average value,  $g$  the gravity, and  $z$  the ver-  
 205 tical direction. The effect of thermal and haline expansion on water density  
 206 can be separated:

$$207 \quad \frac{\partial \rho}{\partial x} = \rho_0 \left( \alpha \frac{\partial \theta}{\partial x} + \beta \frac{\partial S}{\partial x} \right) \quad (5)$$

208 where  $\alpha$  and  $\beta$  are the thermal and haline contraction coefficients of seawater.  
 209

210 The field survey monitored  $\partial\theta/\partial x$  and  $\partial S/\partial x$ , allowing us to estimate the  
 211 expected magnitudes and timing of thermal and salinity currents via scale  
 212 analysis of equations [4](#) and [5](#) (details in [Appendix A.5](#)). Nearshore waters  
 213 exhibited a daily cycle of heating and cooling, and an afternoon salinity drop  
 214 (Fig. 2c). During daytime, warmer and fresher water accumulates near the  
 215 coast, with  $\theta(x)$  and  $S(x)$  decaying exponentially offshore over 100–200 m  
 216 (Fig. [2d](#)). During nighttime cooling, this effect is confined to the nearshore  
 217 region (Fig. [2d](#)). Using the steepest observed gradients, we estimate typical  
 218 current magnitudes of 2 cm/s, 7 cm/s, and 10 cm/s associated to differential  
 219 cooling, heating, and the salinity drop, respectively.

220 The cross-shore diffusion coefficient ( $D$ ) cannot be directly measured from  
 221 field observations. In the surf zone, wave breaking enhances cross-shore mix-  
 222 ing, increasing diffusion compared to the open ocean [\[29\]](#). A rough estimate  
 223 of  $D$  can be derived using the wave breaker height ( $H$ ), wave period ( $\tau$ ),  
 224 and surf zone width ( $X$ ) as  $D = (H \cdot X)/\tau$  [\[30\]](#). For the study site, using  
 225  $H = 0.25$  m,  $\tau = 4$  s, and  $X = 20$  m [\[33\]](#), we estimate  $D = 1.25$   $m^2/s$ , con-  
 226 sistent with reported coastal diffusion values (0.5–1.7  $m^2/s$ ) [\[34\]](#), and tested  
 227  $D = 1.25$   $m^2/s \pm 50\%$  variations in the model.

228 The apparent flow compressibility introduced by the 1D model aligns with the  
 229 behavior of projected incompressible 3D flows, where dimensionality reduc-  
 230 tion inherently relaxes mass conservation in omitted directions. The model  
 231 domain is set to 1 km, with no-flux boundary conditions applied at both ends

232 to ensure conservation of mass. One boundary represents the coastline; the  
 233 other, corresponding to open sea, is located far enough to minimize boundary  
 234 effects. The simulation runs for 7 days to ensure that dynamical equilibrium  
 235 is reached and that the results are independent of initial conditions. We  
 236 tested different values of  $u_\theta$  and  $u_s$  within the range provided by the scaling  
 237 analysis, assuming an offshore exponential decay of the forcing consistently  
 238 with the gradients that drives them (Fig 2d). The numerical analysis was  
 239 developed using the FiPy partial differential equation (PDE) solver [35] and  
 240 is provided in the Appendix A.6.

### 241 3. Results

#### 242 3.1. Coupling between the diel wind pattern and $Chl$

243 Field data show that most biomass was concentrated in a dense nearshore  
 244 stripe a few hundred meters wide. We calculated the total algal biomass  
 245 contained in this stripe,  $Chl_{int}$ , by integrating each  $Chl$  transect within 150  
 246 meters of the coastline. Averaging  $Chl_{int}$  across sampling days revealed a  
 247 clear diurnal pattern: biomass doubled in the afternoon compared to night-  
 248 time and early morning, with the steepest increase from 7:00 AM until a  
 249 sharp decline at 4:00 PM (Fig 2a). In contrast, when  $Chl_{int}$  was calculated  
 250 over the entire transect and averaged daily, total biomass remained fairly  
 251 constant throughout the period ( -4% daily biomass loss, Fig 2a inlet), in-  
 252 dicating no significant net growth or decline of the total population. This  
 253 suggests that the observed daily variation in nearshore biomass is driven by

redistribution rather than biological growth. A comparison of the nearshore  $Chl_{int}$  ( $\leq 150$  m) with cross-shore wind  $W$  revealed a strong temporal correlation between the two timeseries (2a): the onset of morning winds coincided with the steepest biomass increase, while the afternoon wind relaxation coincided with its decline. To explore the time-lagged relationship between the  $Chl_{int}$  and  $W$ , we calculated their cross-correlation as:

$$Cross-Corr(\tau) = \frac{\sum (Chl_{int}(t) - \mu_{Chl_{int}}) \cdot (W(t + \tau) - \mu_W)}{N \cdot \sigma_{Chl_{int}} \cdot \sigma_W} \quad (6)$$

where  $\mu_{Chl_{int}}$  and  $\mu_W$  are the averages of  $Chl_{int}$  and  $W$ ,  $\sigma_{Chl_{int}}$ , and  $\sigma_W$  are the respective standard deviations, and  $N$  the number of points. The cross-correlation pattern reveals a distinct oscillation between positive and negative values, indicating periodic shifts between in-phase and out-of-phase behavior (Fig. 3a). The cross-correlation peaks at  $\tau = 0$ , suggesting that the rise in onshore winds is closely associated with the immediate accumulation of biomass near the coast. The spectral density of the cross-correlation reveals a clear peak corresponding to a 24-hour period (3b), implying that the two time series remain in phase if one is shifted by 24 hours. We performed a sensitivity analysis varying the integration range of  $Chl_{int}$  up to 600 m from the shoreline, and found that synchronization with  $W$  extended to approximately 300 m offshore (see Appendix A.8). Overall, the data suggest that a periodic, wind-driven accumulation mechanism underlies the observed  $Chl_{int}$  pattern.

275 *3.2. Diel retention and dispersion pattern*

276 The cross-shore distribution of *Chl* also exhibited a daily pattern (Fig. 4a).  
277 In the morning, we observed a shoreward exponential increase in algal biomass,  
278 reaching a peak near the shore in the mid-morning, measured at 5 mg/m<sup>3</sup>. In  
279 the afternoon, the *Chl* distribution flattens and the peak is shifted offshore,  
280 relocating between 50 and 100 meters from the coast. This spatial redistri-  
281 bution of the biomass was consistent between different days and suggests the  
282 presence of a mechanism producing a horizontal off-shore displacement of the  
283 biomass.

284 The currents measured in M3 (Fig. 2b) also display a well-defined diurnal  
285 pattern, aligning with the expected diurnal variability of currents driven by  
286 wind advection and cross-shore gradients in temperature and salinity (Fig.  
287 2a, 2c). Onshore currents of  $\sim 1$  cm/s are observed during the night hours  
288 and start increasing at 7:00 AM, following the rise of the onshore winds (Fig.  
289 2b). A sharp flow inversion is observed at 1:00 PM as nearshore currents  
290 start flowing offshore until 6:00 PM. This flow inversion coincides with the  
291 maximum thermal gradient associated with water differential heating. We  
292 can expect this offshore flux to be sustained by the salinity decline, peaking  
293 between 4:00 and 7:00 PM and also contributing to the offshore flux of the  
294 water. Apart from a mild nearshore recirculation cell observed after 6:00 PM,  
295 the ADCP data suggests the absence of a clear 2D structure in the proximity  
296 of the coast. The current direction remains fairly constant along the verti-

cal, suggesting a mainly 1D displacement of the water column governed by horizontal currents.

Fig. 4a shows the comparison between the cross-shore *Chl* transects measured during the field campaign and the profiles of the passive tracer distribution resulting from the 1D model at the corresponding time of the day. The best agreement between model and data is found for thermal and haline currents ranging between 1 cm/s and cm/s, coinciding with the range of currents measured by the current meter in M3 (Fig. 2b). The model result is displayed as an average between the different model runs within this optimal thermal and haline current ranges. Figure 4b shows the diurnal variability of  $Chl_{int}$  calculated at three different distances from shore (100 m, 300 m, 600 m). Overall, the observed diurnal dynamic of the *Chl* field is fairly captured by the 1D approximation, both in terms of the cross-shore spatial distribution (4a) and the integrated profile (4b). The model overall captures the dynamics of the nearshore biomass, the agreement between the model and the data decreasing further away from the coastline.

#### 4. Discussion

In the present study, we adopt a simple advection-diffusion model to analyze the short-term dynamics of the biomass contained in the nearshore coastal stripe, treating *Chl* as a passive, well-mixed tracer. The spatial and temporal resolution considered—sub-kilometer and hourly scales—is rarely

318 addressed in HAB research, which typically focuses on broader spatial and  
 319 longer temporal frameworks. Although this approach deliberately simplifies  
 320 the system by omitting biological variability, it shows high exploratory and  
 321 conceptual value since it allows us to successfully shed light on the roles  
 322 of three interacting physical forcings (wind, temperature and salinity) with-  
 323 out introducing a high level of complexity and parametrization. Field data  
 324 were collected during the decline phase of a small algal outbreak, whose  
 325 multi-day trend (Fig. 2a, inlet) and in-situ measured growth rates [25] sug-  
 326 gest a biologically quiescent regime. It is unlikely that the inclusion in the  
 327 model of processes such as growth, mortality, and grazing would alter our  
 328 results, since they act at larger timescales. Indeed, [36] show that when nutri-  
 329 ent supply or light limitation lacks significant spatial variability, circulation  
 330 and phytoplankton population dynamics are separable: stirring and mixing  
 331 spatially structure plankton distributions, while biological dynamics govern  
 332 population size. Here, measured planktonic metabolism (estimated as net  
 333 community production, NCP per unit *Chl*) was uniform across the transects  
 334 [25], suggesting an underlying spatially homogenous light and nutrient field.  
 335 Given that equation ?? Thus, consistent with [36], we find that cross-shore  
 336 plankton distribution is controlled by physical processes, whereas the magni-  
 337 tude of the variations depends on phytoplankton standing stocks driven by  
 338 nutrient availability, temperature, and biotic interactions (e.g., parasites and  
 339 grazers).

Overall, our approach successfully captures the main dynamics of the bloom and its simplicity allows us to disentangle and isolate key factors. Short-term fluctuations in *Chl* concentration can be explained by periodic particle retention-dispersal mechanisms driven by the opposing flows induced by wind and density gradients (Fig. 3 and 4). The data indicate that changes in the cross-shore wind speed and in *Chl* are synchronized up to 300m from the coastline (Fig. 2a and Appendix A.8). The morning rise of the landwards winds is accompanied by a shoreward exponential increase in biomass. Differential heating and freshwater inputs produce a mass of buoyant water, which is retained near the coast by the onshore winds. Temperature and salinity gradient build up during the day, eventually overcoming the wind, leading to offshore dispersal of phytoplankton in the early afternoon. At nighttime, the nearshore flow inverts, possibly due to differential cooling. The model enable us to isolate individual contributions and probe potential scenarios (see Appendix A.11). Our analysis shows that wind is the primary driver of the overall accumulation pattern, with over a 4-fold increase in the model residuals for 50% variation in wind strength (Fig. ?? and Appendix A.11). Likewise, the weakening of buoyant flows generated by temperature gradients expedite the deviation of diel biomass responses, with significant late afternoon variability.

Thermal and haline currents were estimated via scale analysis, providing order-of-magnitude estimates of the expected flows. The observed biomass

variability was successfully reproduced for a range of values within this estimate, with flows between 1 and 4 cm/s, consistent with ADCP measurements. A nighttime low-magnitude onshore flow ( $\sim 1$  cm/s) was crucial to counteract the afternoon offshore biomass dispersal driven by the thermaline gradient. This onshore flow, although present in the ADCP measurements, cannot be fully explained by the scale analysis, which predicts an onshore cooling-driven flows only after  $\sim 4:00$  AM (see Appendix [Appendix A.5](#)).

The observed physical mechanism of diel retention–dispersion cycles leads to the daily nearshore compression of phytoplankton biomass, potentially intensifying a range of density-dependent processes— such as particle encounters, exudate accumulation, self-shading, and oxygen production or depletion—with significant ecological implications. Increased daytime biomass concentration nearshore can reduce light penetration, limiting growth through self-shading, while also altering oxygen dynamics. The localized accumulation of phytoplankton-derived compounds may create chemical hotspots that influence predator behavior [\[44\]](#) or trigger the emergence from dormancy in parasites [\[45\]](#), and elevated cell densities can increase encounter rates between phytoplankton and their microzooplankton grazers [\[46\]](#), potentially enhancing top-down control and reshaping trophic interactions. In this context, the observed mechanism may not initiate blooms but could contribute to their decline by facilitating predator–prey interactions and amplifying self-regulatory

384 feedbacks, ultimately influencing bloom persistence and termination.

385 Real-world processes surely exhibit a greater degree of complexity than can-  
386 not be captured by this approximation. At other stages of the bloom, the  
387 timescales of biological dynamics may align more closely with those of the  
388 physical redistribution mechanisms considered here. During the exponential  
389 growth phase, diel gradients in light, nutrients, and temperature can drive  
390 differential growth rates between nearshore and offshore waters, overlapping  
391 with the proposed daily redistribution dynamic. Similarly, in the termina-  
392 tion phase, severe grazing or parasitism might lead to a sharp collapse of the  
393 population. Behavioral factors influencing the diurnal vertical positioning of  
394 cells within the water column may inhibit their dispersion by currents and  
395 alter their horizontal transport [18, 37, 38], rendering the one-dimensional  
396 approximation overly simplistic. Likewise, nearshore flows likely has two-  
397 and three-dimensional structures. The flow produced by differential heating  
398 and cooling is intrinsically not symmetric [32], with water heating stratifying  
399 a superficial buoyant plume, while water cooling produces a more turbulent  
400 vertically mixed flow. Although the measured nearshore currents do not  
401 show clear 2D structures, offshore moored data suggest that the transitional  
402 flow between the onshore and offshore circulation patterns is characterized  
403 by the formation of two counterrotating convection cells (Fig. 5). This  
404 pattern may partially arise from unresolved spatial structures in buoyancy  
405 forcing, driven by the collinear interaction between thermal and haline flows

406 (see Appendix [Appendix A.9](#)). Such complexity likely modulates water  
407 residence times and affects the timing of biomass redistribution along the  
408 transect, contributing to the model–data mismatch observed offshore. These  
409 2D structures cannot be fully characterized with the deployed instrumenta-  
410 tion and not included in the model. Other bio-physical processes might also  
411 contribute to coastal plankton accumulation, such as the amplification of the  
412 thermally driven circulation by the bloom itself through increased water light  
413 absorption [\[39, 40\]](#).

414 Despite these limitations, the model offers valuable insights into the key  
415 mechanisms driving phytoplankton variability and provides a framework for  
416 understanding its transport in the nearshore zone, a key factor in coastal  
417 ecosystem management. The results presented herein reveal how competing  
418 physical forces drive the accumulation/dispersion of nearshore phytoplank-  
419 ton. The model’s ability to reproduce observed patterns underscores its  
420 value in interpreting data of beach water quality, and its simplicity allows  
421 for application to other passive tracers dispersed in nearshore waters such as  
422 municipal discharges, microplastics, or the distribution and fate of eggs and  
423 larvae from littoral organisms.

## 424 **Acknowledgments**

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#### 434 **Author contributions**

435 **G. Basterretxea:** Conceptualization, Funding acquisition, Methodology,  
436 Writing – review and editing. **I. Tuvál:** Conceptualization, Funding acqui-  
437 sition, Methodology, Supervision, Writing – review and editing. **M. Zanoli:**  
438 Methodology, Formal analysis, Writing – original draft.

| Variable names, symbols, units |                                                   |
|--------------------------------|---------------------------------------------------|
| $Chl$                          | Chlorophyll [ $mg/m^3$ ]                          |
| $Chl_{int}$                    | Integrated Chlorophyll [ $mg$ ]                   |
| $x$                            | Cross-shore direction [ $m$ ]                     |
| $t$                            | Time [ $s$ ]                                      |
| $D$                            | Cross-shore diffusion coefficient [ $m^2/s$ ]     |
| $\theta$                       | Seawater temperature [ $^{\circ}C$ ]              |
| $S$                            | Seawater salinity [PSU]                           |
| $W$                            | Cross-shore wind [ $m/s$ ]                        |
| $u$                            | Total cross-shore advection current [ $m/s$ ]     |
| $u_s$                          | Salinity driven advection current [ $m/s$ ]       |
| $u_{\theta}$                   | Temperature driven advection current [ $m/s$ ]    |
| $u_W$                          | Wind driven advection current [ $m/s$ ]           |
| $\alpha$                       | Seawater thermal expansion coefficient [ $1/K$ ]  |
| $\beta$                        | Seawater haline expansion coefficient [ $1/PSU$ ] |
| $\tau$                         | Wave period [ $1/s$ ]                             |
| $H$                            | Wave breaker height [ $m$ ]                       |
| $X$                            | Surf zone width [ $m$ ]                           |
| $\rho$                         | Water density [ $kg/m^3$ ]                        |
| $\rho_0$                       | Average water density [ $kg/m^3$ ]                |
| $z$                            | Water depth [ $m$ ]                               |
| $g$                            | Gravitational constant [ $m/s^2$ ]                |

Table 1: List of used abbreviations, symbols and units

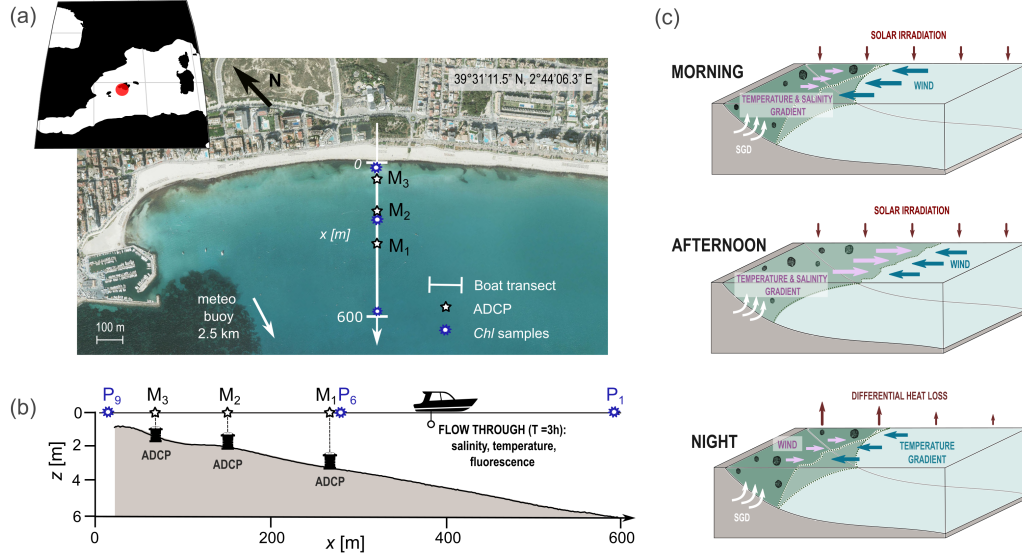


Figure 1: **(a)**: Sampling site in Palma Beach, Mallorca, Spain. The cross-shore transect travelled by the sampling boat is shown in white and sets the  $x$ -coordinate of the analysis. The position of the Acoustic Doppler Current Profiler (ADCP) moorings (M1, M2, M3) and the *Chl* sampling points (P1, P6, P9) are marked on the transect line. **(b)**: Beach bathymetry along the transect **(c)**: Schematics of the natural forcings acting on the system at different times of the day. During daytime, the onshore sea-breeze forcing competes with an off-shore buoyant flow driven by the differential heating of the water and coastal SGD. During the night, the winds weaken and reverse, while differential cooling of the water drives an on-shore flow.

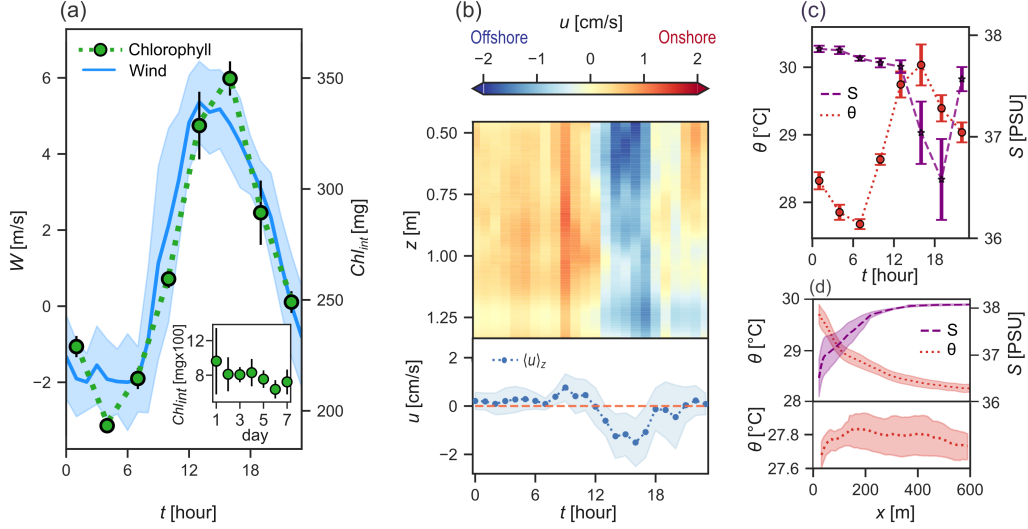


Figure 2: **(a)**: Hourly averages of cross-shore winds and total biomass accumulated near the shoreline  $Chl_{int}$  ( $< 150$  m) during the 6-day sampling period (excluding data from storm day on 22/07/2018).  $Chl_{int}$  [mg] is obtained by integrating  $Chl$  concentration profiles [ $\text{mg}/\text{m}^3$ ] along the transect direction and depth profile, assuming a homogeneous distribution of  $Chl$  along the water column. Inset: daily  $Chl_{int}$  average calculated over the entire transect ( $< 600$  m). **(b)** Typical daily patterns of the cross-shore currents from ADCP data measured at M3 (top), and its vertical average along  $z$  (bottom). The typical current for each hour is calculated by averaging the currents measured at that specific hour across all days of the sampling period. The entire water column flows cohesively in one direction with no clear recirculation cells developing near the shore, suggesting a mainly 1D flow. Currents shown in the bottom panel are obtained as the vertical average of those shown in the top panel. **(c)** Typical values of water temperature and salinity near the shoreline (25 m) at different hours of the day. The typical temperature and salinity for each hour is calculated by averaging the temperature and salinity measured at that specific hour across all days of the sampling period. **(d)** Temperature and salinity profiles from the shoreline to 600 m at the times showing the strongest cross-shore gradients. Up: temperature profile associated to differential heating at 1:00 PM (red dotted line), salinity profile during the salinity drop at 19:00 PM (dashed violet line). Down: temperature profile associated to differential cooling at 07:00 AM.

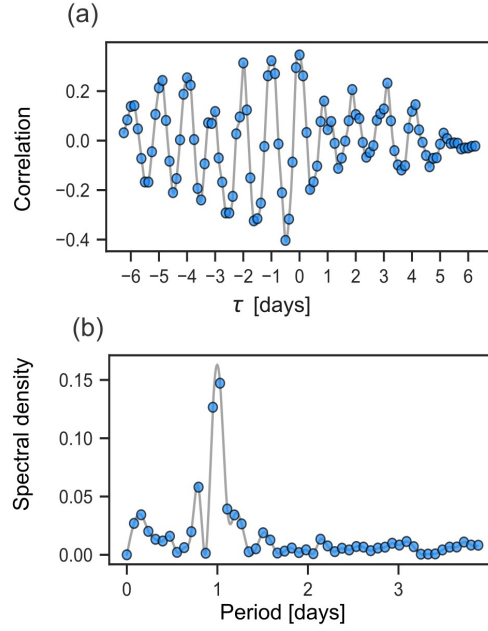


Figure 3: **(a)** Cross-Correlation between Cross-Shore Wind and  $Chl_{int}$  (<150 m) Time-series. The cross-correlation is calculated on the mean-subtracted timeseries. The maximum correlation is found at zero lag, and successive maximum positive values are observed every 24 hours, suggesting that variations in nearshore  $Chl$  are synchronized with variations in the cross-shore wind. **(b)** Spectral density of the Fourier transform of the cross-correlation in Panel B. A clear peak at a 1-day periodicity is observed.

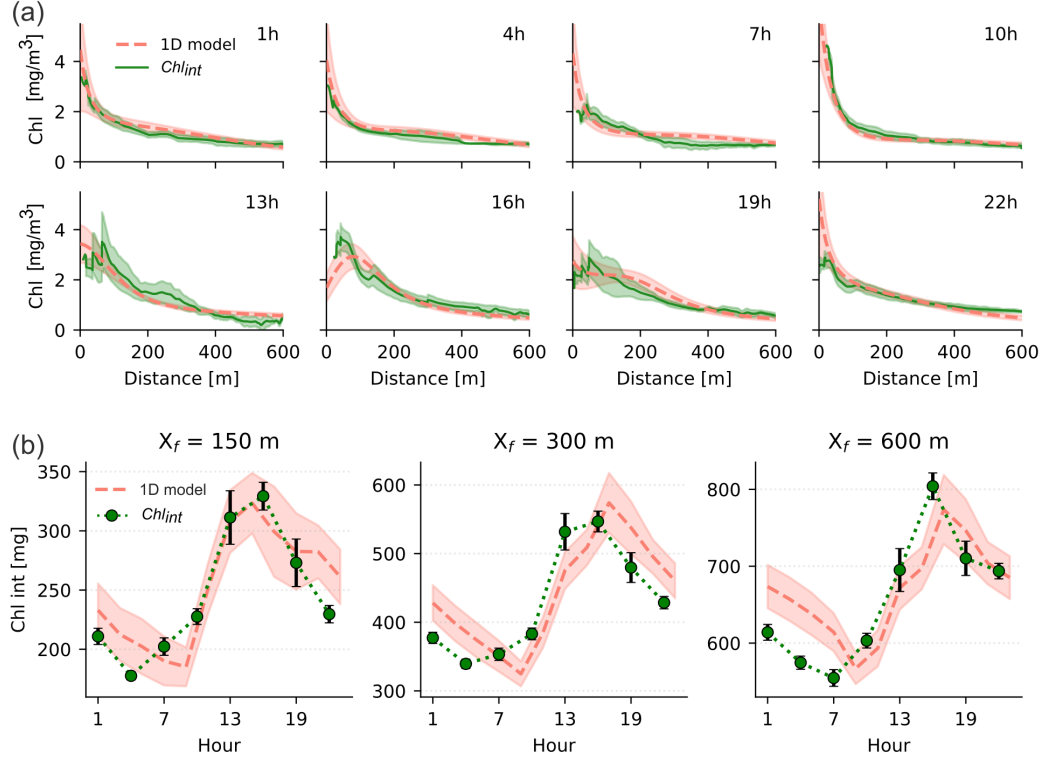


Figure 4: **(a)** Cross-shore transects of  $Chl$  biomass [ $\text{mg}/\text{m}^3$ ], averaged at the same hour across all days of the sampling period (green line). The pink line shows the biomass accumulation predicted by the 1D model as a result of the dynamical equilibrium between the forcings resulting from the wind and the thermal-haline syphon. The model output is represented as the ensemble mean of all simulations across the range of diffusivities and advection flows explored (dashed line), restricting the ensemble to a thermal heating flow  $u_\theta > -4$  cm/s. The number in parentheses represents the average daily residual between observed and modeled  $Chl$  concentrations at matching transect locations for the best-performing model run, expressed in  $\text{mg}/\text{m}^3$  **(b)**  $Chl_{int}$  of a typical day at 150m, 300m and 600m from the coast (green dots) obtained integrating the  $Chl$  profiles in panel a. The pink dashed line marks the model ensemble mean, and the shaded area its standard deviation.

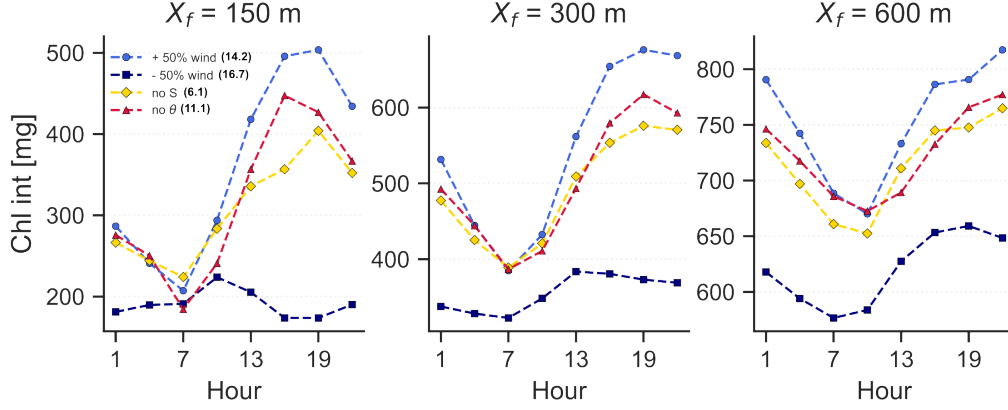


Figure 5: Integrated profiles of average cross-shore transects of *Chl* biomass [mg] up to three different distances from the coast (150, 300 and 600 m), compared with that modelled in 4 alternative scenarios: +50% and -50% wind, no salinity-driven flow and no temperature-driven flow. The number in parentheses represents the average daily residual between observed and modeled chlorophyll concentrations at matching transect locations of each model run, expressed in mg *Chl*/m<sup>3</sup>

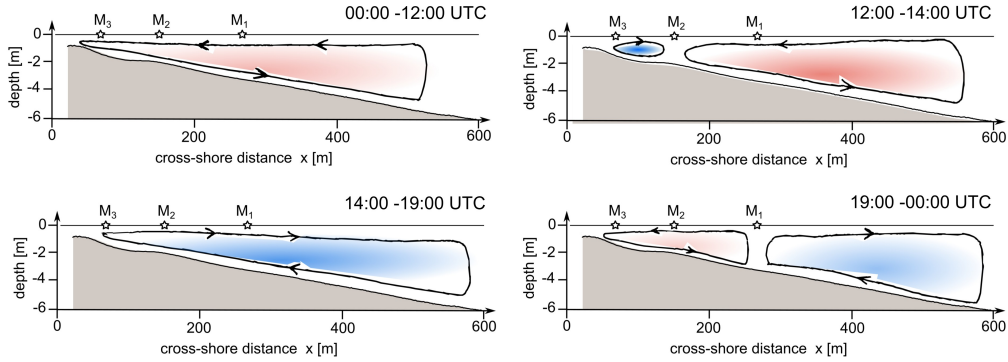


Figure 6: Current data from moored instrumentation in M1,M2,M3 suggests that the transitions between the morning onshore flow and the afternoon offshore flow are characterised by the formation of two counter-rotating convection cells. In both cases, the transition flow starts close the coastline and gradually extends seawards. The onshore morning flow, driven by winds and differential cooling of the water, is counteracted by the kick in of a thermo-haline syphon, the buoyant water flowing offshore. After sunset, differential cooling of the littoral waters prompts the shorewards flow.

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## 593 Appendix A. Appendix

### 594 Appendix A.1. Transects calibration: from fluorescence to *Chl* concentra- 595 tion

596 Water fluorescence was measured using an Enviro-T in-line fluorometer along  
597 the transects shown in Fig. 1 of the manuscript, serving as a proxy for mi-  
598 croalgae biomass as described in 4. For each transect, water samples were  
599 collected at three points (P1, P6, and P9 in Fig. 1), and *Chl* concentration  
600 was measured by filtration.

601 Detritus and debris, such as suspended leaf fragments, can become lodged in  
602 the fluorometer and interfere with sensor readings, causing anomalies—peaks  
603 if the debris is fluorescent or valleys if it is non-fluorescent. To remove these  
604 artifacts, all fluorescence data were post-processed using a Python routine.  
605 Major peaks and valleys in the signal were identified, fitted to a Gaussian  
606 model, and removed. The gap in the signal was filled by linear interpo-  
607 lation and the reconstructed signal was smoothed with a third-order Sav-  
608 itzky–Golay polynomial filter, which reduces noise while preserving the pri-  
609 mary signal pattern.

610 Fluorescence readings were converted to *Chl* concentrations using water sam-  
611 ples for calibration. Since the fluorescence signal depended on instrument  
612 conditions at the time of measurement (e.g., battery level), a three-point cal-  
613 ibration was performed for each transect individually. A linear relationship  
614 was assumed between the instantaneous fluorescence (*Flu*) and *Chl* concen-

615 tration. For a transect measured at time  $t_0$ :

$$616 \quad Flu(x, t_0) = a(t_0) \cdot Chl(x, t_0) + b(t_0) \quad (A.1)$$

617 where  $a(t)$  and  $b(t)$  are specific to each individual transect and are obtained  
 618 performing a linear fit between the  $Chl$  point values in P1,P6,P9 and the  
 619 corresponing measured values of  $Flu$  at the same locations. Fig. [A.7](#) il-  
 620 lustrates the agreement between point chlorophyll measurements and their  
 621 corresponding calibrated estimates across three sampling stations.

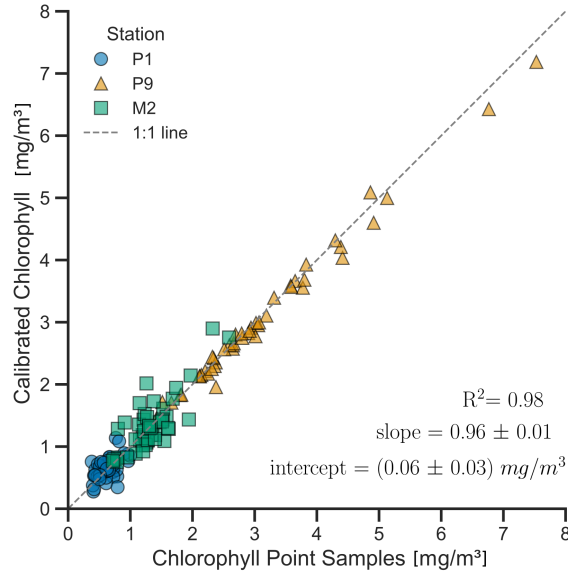


Figure A.7: Comparison between point chlorophyll measurements ( $\text{mgm}^{-3}$ ) and corresponding chlorophyll estimates from the calibration. Data from the three stations are shown: P1 (circles), P9 (triangles), and M2 (squares). The dashed line indicates the 1:1 relationship. Close alignment of points with the line suggests good agreement between observed and calibrated values across all stations.

622 *Appendix A.2. CTD vertical profiles*

623 To assess the vertical structure of the water column during sampling, CTD  
 624 profiles of temperature, salinity, and chlorophyll fluorescence were collected  
 625 at multiple stations on 20 July 2018 (Fig. A.8). These profiles reveal verti-  
 626 cally homogeneous distributions across all measured parameters within the  
 627 0–7 m depth range, indicating well-mixed conditions throughout the sampled  
 628 layer. While data from the uppermost 0.5 m of the water column are unavail-  
 629 able—where surface accumulation of larger cells could potentially occur—the  
 630 lack of thermal or haline stratification below this skin layer, combined with  
 631 the absence of subsurface chlorophyll maxima or phytoplankton aggrega-  
 632 tions, further supports the inference of vertical homogeneity in these shallow  
 633 coastal waters.

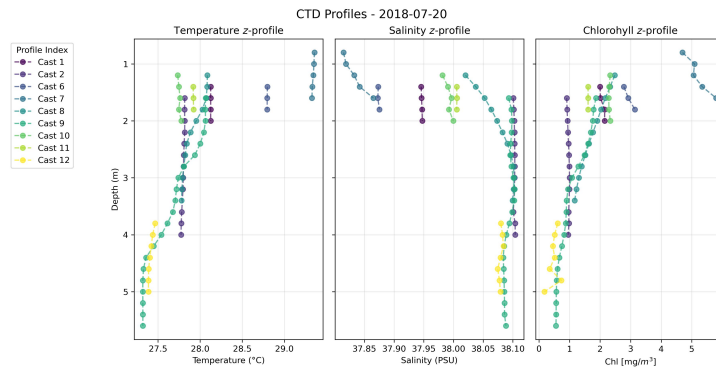


Figure A.8: CTD vertical profiles collected on 20 July 2018 at multiple stations along the transects. Each panel shows measurements as a function of depth of: (left) temperature ( $^{\circ}\text{C}$ ), (middle) salinity (PSU), and (right) chlorophyll-a concentration ( $\text{mg}/\text{m}^3$ ).

634 *Appendix A.3. ACDP profiles in M1, M2, M3*

635 Figure A.9 displays the currents measured by the ACDP during the field  
636 survey at three transect locations (M1, M2, and M3), placed at 70 m, 150  
637 m, 280 m from the coastline. For each hour of the day, the current vertical  
638 profile is calculated by averaging the currents measured at that specific hour  
639 across all days of the sampling period. These profiles, therefore, represent the  
640 typical daily current patterns at the study site. Overall, the current direction  
641 appear consistent along the vertical direction down to a depth of 1–1.5 meters,  
642 indicating that the flow is predominantly one-dimensional within this layer.  
643 Notably, the offshore flow observed at M3 in the early afternoon (between  
644 1 PM and 7 PM) weakens significantly at M2 and re-emerges at M1 with  
645 a two-hour delay (between 3 PM and 9 PM). The nighttime onshore flow,  
646 which is weaker at M3, reaches speeds of up to 2 cm/s at M2 and M1. This  
647 onshore flow can only be partially attributed to the differential cooling of  
648 nearshore waters compared to the open sea and is introduced into the model  
649 as a phenomenological current. Although the mechanisms driving this flow  
650 are not fully explained by the "thermal siphon" effect, the model suggests  
651 that the presence of this flow is essential to counterbalance the afternoon  
652 flushing of biomass and accurately reproduce the observed daily biomass  
653 pattern.

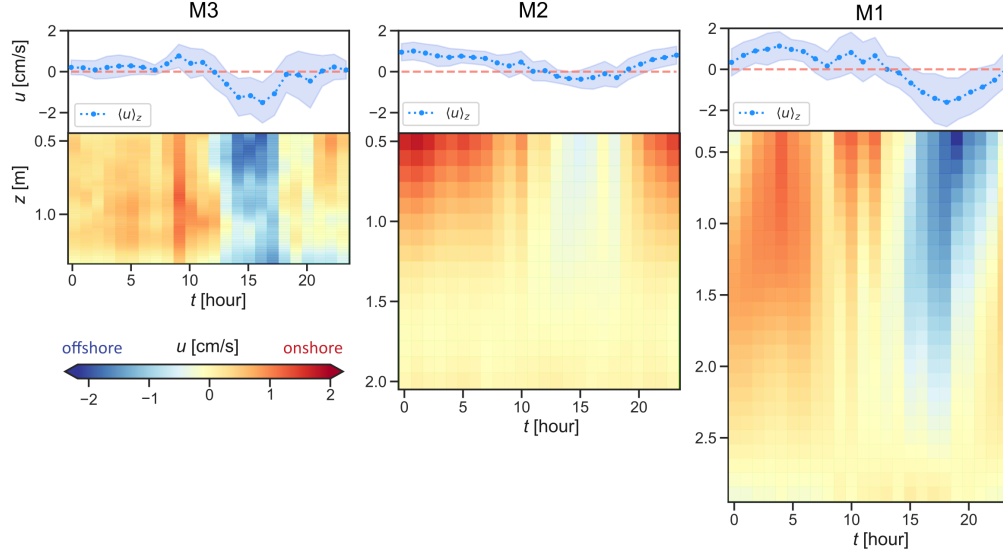


Figure A.9: ADCP data in M3 (70 m), M2 (150 m), M1 (280 m) measured during the field survey. Each colormap represents the vertical profile of the flow obtained averaging the current data measured at the same hour of the day across different days. The upper plots represent the vertical average of the flows depicted in the colormaps. Onshore flows are represented in red and offshore flows in blue.

#### 654 *Appendix A.4. Tidal contribution to the advective term*

655 Tidal forcing was not explicitly isolated in our analysis. However, tidal con-  
 656 tributions — though relatively weak in magnitude (0.7–1.3 cm/s)— could  
 657 compete with the other forcings, particularly during nighttime hours when  
 658 wind diminishes. To estimate the potential influence of tides on the pro-  
 659 posed retention–dispersion mechanism, we compared sea level data recorded  
 660 at station M3, the closest point to the shoreline, with ADCP measurements  
 661 of total currents, which integrate all physical forcings. The left panel of Fig.  
 662 A.10 shows sea level (SL) at M3 during the field survey, while the right panel

663 presents an hourly averaged SL profile constructed in the same manner as  
 664 the "typical day" used for other model variables (e.g., wind, temperature,  
 665 salinity). Although tidal cycles do not follow a strict 24-hour cycle, this av-  
 666 eraging provides a first-order approximation of their diel influence. Notably,  
 667 the tidal signal appears roughly in anti-phase with the observed diel current  
 668 patterns: ebb tide (falling sea level in the morning) would typically drive off-  
 669 shore flow, and flood tide (rising sea level in the afternoon) would promote  
 670 onshore flow. In contrast, ACDP measurements show the reverse —onshore  
 671 flow in the morning and offshore flow in the afternoon—indicating that the  
 672 diel current signal is not primarily tidal in origin. It is likely, instead, that  
 673 tidal currents partially dampen the net circulation driven by diel forcing.  
 674 This suggests that the retention–dispersion dynamics described in the main  
 675 text may be even more pronounced in the absence of tidal modulation.

#### 676 *Appendix A.5. Scale analysis of the temperature and salinity driven cur-* 677 *rents*

678 The advective currents generated by density gradients, which arise from  
 679 cross-shore variations in temperature and salinity, are estimated through  
 680 scale analysis. This approach provides the expected order of magnitude for  
 681 the flows driven by these gradients. If we consider a fluid system initially  
 682 in an equilibrium flow to which a reversed temperature gradient is suddenly  
 683 applied, the system will try to restore stability through a convective mo-  
 684 tion. This stabilizing convection is governed by the unsteady inertia of the

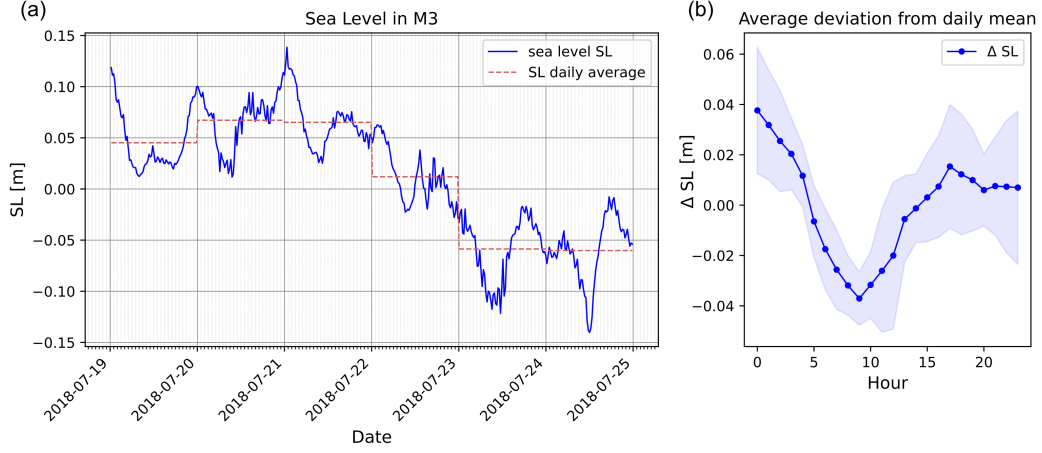


Figure A.10: (a): Sea level (SL) time series during the field survey as measured in M3. The red dotted line is the daily mean of the sea level. (b) Average hourly sea level deviation relative to daily mean (SL). For each day, SL was calculated relative to that day's mean value. These hourly deviations were then averaged across all days to represent the typical daily cycle of sea level fluctuations.

685 water:

$$686 \quad \rho_0 \frac{\partial^2 u}{\partial t \partial z} = g \cdot \frac{\partial \rho}{\partial x} \quad (\text{A.2})$$

687 where  $u$  is the cross-shore current,  $\rho_0$  is the average density of the water,  
 688  $g$  the gravity, and  $x$ ,  $z$  and  $t$  are the cross-shore horizontal direction, the  
 689 vertical direction, and time respectively. We assume that density gradient  
 690 due to thermal expansion and the one due to haline expansion can be sepa-  
 691 rated:

$$692 \quad \frac{\partial \rho}{\partial x} = \rho_0 \left( \alpha \frac{\partial \theta}{\partial x} + \beta \frac{\partial S}{\partial x} \right) \quad (\text{A.3})$$

693 where  $\alpha$  is the thermal contraction coefficient of seawater, and  $\beta$  the haline  
 694 contraction coefficient.

695 Following Monismith 1990 [32], the scale analysis on equations A.2 and A.3  
 696 provides an estimate of the expected magnitudes for the thermal and salinity  
 697 currents  $U_\theta$  and  $U_S$ , together with the typical spin-up timescales  $T_\theta$  and  $T_S$   
 698 for such currents to develop from an initial state of rest:

$$699 \quad U_\theta = \sqrt{g \cdot \alpha \cdot \Delta\theta \cdot L_z^\theta} \quad U_S = \sqrt{g \cdot \beta \cdot \Delta S \cdot L_z^S} \quad (\text{A.4})$$

700 and:

$$701 \quad T_\theta = \frac{U_\theta \cdot L_x^\theta}{g \cdot \alpha \cdot \Delta\theta \cdot L_z} \quad T_S = \frac{U_S \cdot L_x^S}{g \cdot \beta \cdot \Delta S \cdot L_z} \quad (\text{A.5})$$

702 where  $L_x^\theta$  and  $L_x^S$  are the typical horizontal scales of the temperature and  
 703 salinity gradients, and  $L_z$  is the typical scale of the system in the vertical  
 704 direction.

705 Typical values for the steepness of the gradients  $\Delta\theta/L_x^\theta$  and  $\Delta S/L_x^S$  are esti-  
 706 mated from the temperature and salinity fields  $\theta(x, t)$  and  $S(x, t)$  measured  
 707 during the field campaign (figure 2, panel B). We use  $L_z = 1$  m as vertical  
 708 scale of the system. This vertical estimate is extracted from the ACDP cur-  
 709 rent profiles, which show that the currents flow cohesively within the first  
 710 1 m layer (Fig. A.9). As horizontal scales, we use  $L_x = 600$  m for flows  
 711 driven by thermal heating and salinity drop. For thermal cooling, we use  $L_x$   
 712  $= 200$  m, as the morning thermal inversion does not extend throughout the  
 713 transect, but is restricted to the nearshore band (Fig. A.13a-b). We consider  
 714 a typical temperature variation of  $\Delta\theta = 1.5$  °C for differential heating,  $\Delta\theta =$

0.1 °C for differential cooling, and  $\Delta S = 1$  PSU for the salinity drop. This parameters result in a thermal afternoon offshore current of  $U_\theta = 7$  cm/s with spin up time  $T_\theta = 3$  h, in a haline current of  $U_S = 10$  cm/s with spin up time  $T_S = 1.5$  h and a morning onshore thermal current of  $U_\theta = 2$  cm/s with spin up time  $T_\theta = 3$  h.

In our approximation, we assume the buoyancy flows to act on time windows centered in the moments of steepest temperature and salinity gradients (refer to Fig. A.13b-d). We assume the steepest gradients to appear at 7:00 AM (differential cooling), 2:30 PM (differential heating) and 5:30 PM (salinity drop). For the latter two, the chosen time corresponds to the central value between the sampling times when maximum temperature and salinity gradients where measured.

We assume that the system required a time  $T_\theta$  and  $T_S$  to reach the flows  $U_\theta$  and  $U_S$  at 7:00 AM, 2:30 PM and 5:30 PM for the flows driven by differential cooling, differential heating and the salinity drop respectively. We assume that, symmetrically, the "spin-down" time for the system to stop flowing after the disappearance of the density gradient will be approximately the same as the timescale of the spin-up time, as we expect the slowing down of the system to be governed by the same physical processes as the spin-up. We thus define the buoyancy currents to act within a time window twice the typical spin-up timescale, centered around the moments of maximum gradients. This corresponds to differential cooling acting between 4 AM and

737 10 AM, differential heating between 11:30 AM and 5:30 PM, and the salinity  
738 forcing to act between 4:00 PM and 6:30 PM.

739 The flows predicted by the scale analysis overall match in terms of timing  
740 and directions with the ones measured by the ACDP at the location closest  
741 to the shore (M3). The scaling predicts an offshore flow driven by buoyancy  
742 flows between 11:30 AM and 6:30 PM, coinciding with the measured flows  
743 (see Fig. [A.9](#)), partially counterbalanced by the onshore wind-driven flows.  
744 The onset of the winds at 7:00 AM together with the action of flows driven by  
745 differential cooling between 4 AM and 10 AM matches well with the increase  
746 of the onshore flow measured in M3 at 7:00 AM. The scale analysis cannot  
747 explain the onshore flow measured throughout the full night hours, which  
748 is introduced in the model as a phenomenological current between 9:00 PM  
749 and 4:00 AM.

#### 750 *Appendix A.6. Codes of the 1D model*

751 The 1D advection-diffusion equation is implemented using the FiPy partial  
752 differential equation (PDE) solver. The code developed is available at the git  
753 repository [https://gitlab.com/medeazanoli/1d\\_advection\\_diffusion\\_](https://gitlab.com/medeazanoli/1d_advection_diffusion_wind_thermo-haline_syphon.git)  
754 [wind\\_thermo-haline\\_syphon.git](https://gitlab.com/medeazanoli/1d_advection_diffusion_wind_thermo-haline_syphon.git)

#### 755 *Appendix A.7. Bloom decline*

756 We calculated the total nearshore chlorophyll by integrating the measured  
757 *Chl* field across the entire transect (600 m). Figure [A.11](#) shows the temporal

758 evolution of total algal biomass throughout the sampling period. The data  
 759 suggest that the system was near saturation, with biomass entering a phase of  
 760 gradual decline. A linear fit to the data indicates that total biomass decreased  
 761 at a rate of  $-36 \pm 11 \text{ mg/day}^{-1}$ , corresponding to an average daily loss of  
 762 approximately 4% relative to the initial chlorophyll concentration.

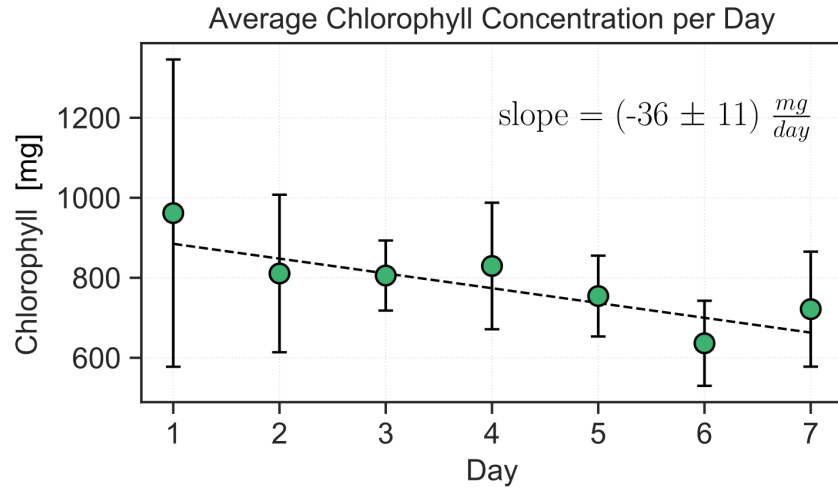


Figure A.11: Mean chlorophyll contained in the transect (600 m) measured over seven days. The dashed black line denotes the best linear fit to the data, with slope  $(-36 \pm 11 \text{ mg/day}^{-1})$ .

763 *Appendix A.8. Sensitivity analysis of the cross correlation between Chl and*  
 764 *wind*

765 In section [3.1](#) of the Results, we explored the time-lagged relationship be-  
 766 tween the cross-shore winds and the *Chl* biomass accumulated in the nearshore  
 767 stripe, and found that the two time-series are synchronized between each  
 768 other. The nearshore stripe was defined to extend between  $X_i = 25 \text{ m}$  and  
 769  $X_f = 150 \text{ m}$ , where  $[X_i, X_f]$  represents the integration interval. The lower

770 limit  $X_i = 25$  m is set by a practical constraint of the sampling, as the vessel  
 771 could not approach closer to the coastline. Thus, 25 m is the closest distance  
 772 at which there is enough fluorescence data to calculate a *Chl* average for each  
 773 transect. To check that the cross-correlation result is robust, we varied the  
 774 value of  $X_f$  between 80 m and 590 m and recalculated the cross correlation  
 775 for each value of  $X_f$ . Fig. A.12a shows the cross-correlation for different  
 776 values of  $X_f$ , and A.12b the time lag  $\tau$  in equation 6 of the manuscript re-  
 777 sulting in the maximum correlation. We find that the nearshore biomass  
 778 stripe is homogeneously synchronized with the cross-shore winds up to  $X_f$   
 779 = 310 m.

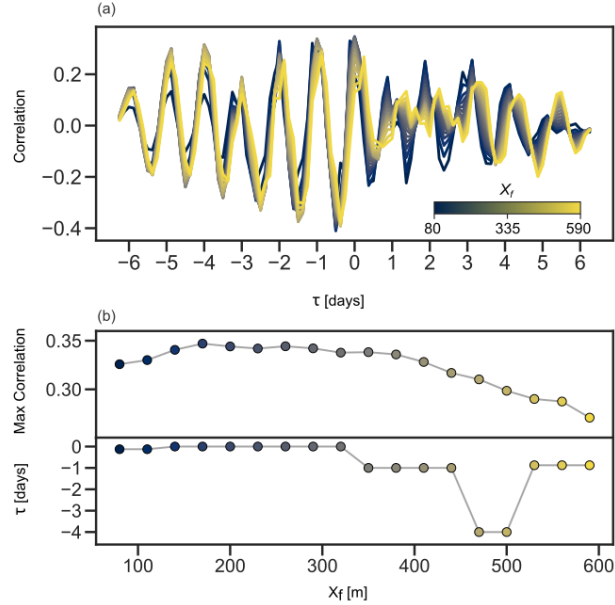


Figure A.12: **(a)** Cross-Correlation between Cross-Shore Wind and *Chl* Timeseries for different values of the biomass integration range  $X_f$ , from  $X_f = 80$  m (dark blue line) to  $X_f = 590$  m (yellow line) **(b)** The maximum correlation obtained for each tested value of  $X_f$  (up), and the corresponding lag  $\tau$  in days (down).

780 *Appendix A.9. Temperature and salinity cross-shore transects*

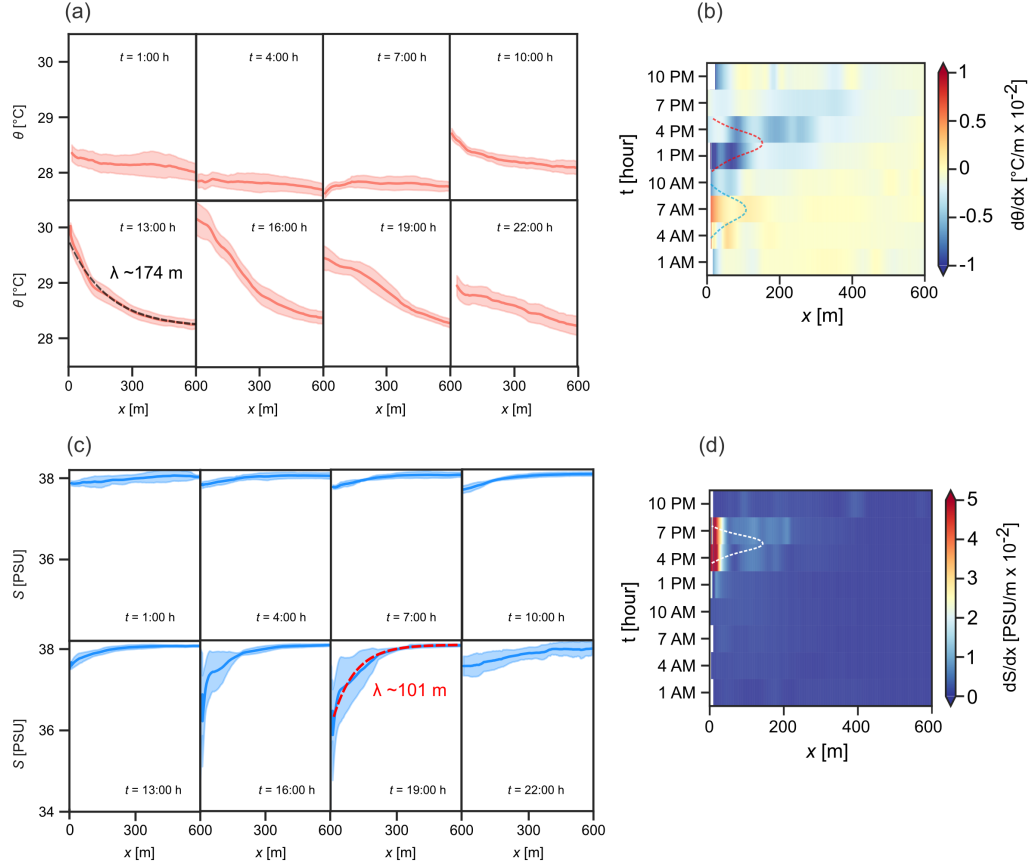


Figure A.13: Cross-shore transects of temperature (a) and salinity (b), averaged across different sampling days. The dashed lines represent the best fit to an exponential function for the temperature profile at 1:00 PM and the salinity profile at 7:00 PM, when the steepest gradients are observed. The reported value of  $\lambda$  is the constant of the exponential fit. The colormaps on the right show the cross-shore temperature (b) and salinity (d) gradients across the transects shown in (a) and (c) respectively. The derivative in  $x$  is calculated from the spline interpolation of each transect. The dotted Gaussian lines indicate the time windows during which buoyancy-driven flows are introduced into the 1D model (in (b): differential cooling marked by the blue line centered at 7:00 AM; differential heating by the red line centred at 2:30 PM. in (d): salinity drop marked by the white line centered at 5:30 PM).

781 Fig. A.13 shows the temperature and salinity cross-shore transects measured  
 782 during the field campaign. The transects in Fig. A.13a-c represent the

783 typical daily cycle of temperature and salinity in the nearshore zone. These  
784 are obtained by averaging the transects measured at the same hour across  
785 different days of the field survey. The daily evolution of temperature and  
786 salinity gradients is consistent across the sampling days, revealing a clear  
787 diurnal cycle in the nearshore waters.

788 The temperature difference between the nearshore waters and the open sea  
789 reaches approximately 2 °C during the day, with the steepest gradient ob-  
790 served at 1 PM. At this time, the temperature profile increases exponentially  
791 toward the shore, characterized by an exponential decay constant of  $\lambda \sim$   
792 174 m. In the afternoon, the temperature gradient relaxes as the nearshore  
793 waters cool more rapidly, and the steepest gradient shifts offshore. By the  
794 end of the cooling phase, around 7:00 AM, a mild thermal inversion occurs,  
795 with the nearshore waters being approximately 0.2 °C colder than the open  
796 sea.

797 The salinity cycle is similarly consistent across sampling days. A sharp salin-  
798 ity drop of 2 PSU is observed between 4 PM and 7 PM, when salinity de-  
799 creases exponentially towards the shore ( $\lambda \sim 101$  m). Fig. A.13b-d shows  
800 the cross-shore gradients of temperature and salinity obtained via spline in-  
801 terpolation of the profiles in Fig. A.13a-c. The dotted lines indicate the time  
802 windows during which buoyancy-driven currents are implemented in the 1D  
803 model. In the model, differential cooling drives an onshore flow from 4:00 to  
804 10:00 AM, peaking around the maximum cooling at 7:00 AM. The differen-

805 tial heating window is centered at 2:30 PM, spanning from 11:30 AM to 5:30  
806 PM. The salinity-driven flow is active between 4:00 PM and 7:00 PM.

807 In the model, thermal and haline buoyancy flows are treated independently.  
808 However, their overlapping time windows can lead to collinear interactions,  
809 introducing more complex patterns. The onset of the haline flow may flush  
810 warmer waters offshore with potentially strong velocities (up to 10 cm/s,  
811 based on a scale analysis of the steepest observed haline gradient), but its  
812 influence remains spatially confined ( $\lambda \sim 100$  m; Fig. A.13). The combined  
813 effect of these collinear flows could advect warmer waters seaward, displacing  
814 the peak thermal gradient offshore. This partial dependency between buoy-  
815 ancy forcings may contribute to the late-afternoon recirculation cell observed  
816 in Fig. 6: onshore wind forcing—strongest where land–sea temperature gra-  
817 dients are sharpest—may counteract offshore buoyancy flows nearshore. Be-  
818 yond  $\sim 200$  m, the thermal flow sustained by the warm water offshore ad-  
819 vection could dominate, generating a two-dimensional flow structure that  
820 modulates water residence times and influences the timing of biomass redis-  
821 tribution across the transect.

#### 822 *Appendix A.10. Sensitivity analysis of the 1D model*

823 The 1D advection-diffusion model was run for different magnitudes of ad-  
824 vective currents associated to the thermal ( $u_\theta$ ) and haline flows ( $u_S$ ), with  
825 the addition of a nighttime onshore phenomenological current  $u^*$  which is  
826 observed in the ACDP data. For the morning flow driven by differential cool-

ing, we explored current values between 1  $cm/s$  and 4  $cm/s$  between 4:00  
 AM and 10:00 AM. For the differential heating and salinity-driven flows, we  
 explored current values between 1  $cm/s$  and 8  $cm/s$ , between 11:30 AM and  
 5:30 PM, and between 4:00 PM and 6:30 PM respectively. The presence  
 of an additional phenomenological onshore current  $u^*$  is explored after  $u_S$   
 dissipates (between 9:00 PM and 4:00 AM). A range of magnitudes for  $u^*$   
 is explored between 0  $cm/s$  (no current) and 4  $cm/s$  as for the flow driven  
 by differential cooling. The model was run with all combinations of these  
 parameters with the three explored values of cross-shore diffusivities  $D =$   
 1.25  $m^2/s$ ,  $D = 0.675 m^2/s$  and  $D = 1.875 m^2/s$ . For each model run, we  
 calculate the residuals between the biomass data and the distribution of the  
 modelled passive tracer along each individual transect. Averaging the resid-  
 uals across all eight daily transects gives a single mean value of residual for  
 each model run. Fig. A.14 shows the value of the mean residual for each of  
 the explored parameters combination.

Finally, we explored the following combinations of advective flows:

- $u_S$  : -1  $cm/s$ , -2,  $cm/s$ , -4,  $cm/s$ , -8  $cm/s$  ( $\lambda = 120$  m)
- $u_{\theta}^{onshore}$  : 1  $cm/s$ , 2,  $cm/s$ , 4 ( $\lambda = 100$  m)
- $u_{\theta}^{offshore}$  : -1  $cm/s$ , -2,  $cm/s$ , -3  $cm/s$ , -4,  $cm/s$ , -8  $cm/s$  ( $\lambda = 200$  m)
- $u^*$  : 1  $cm/s$ , 2,  $cm/s$ , 4  $cm/s$  ( $\lambda = 100$  m)

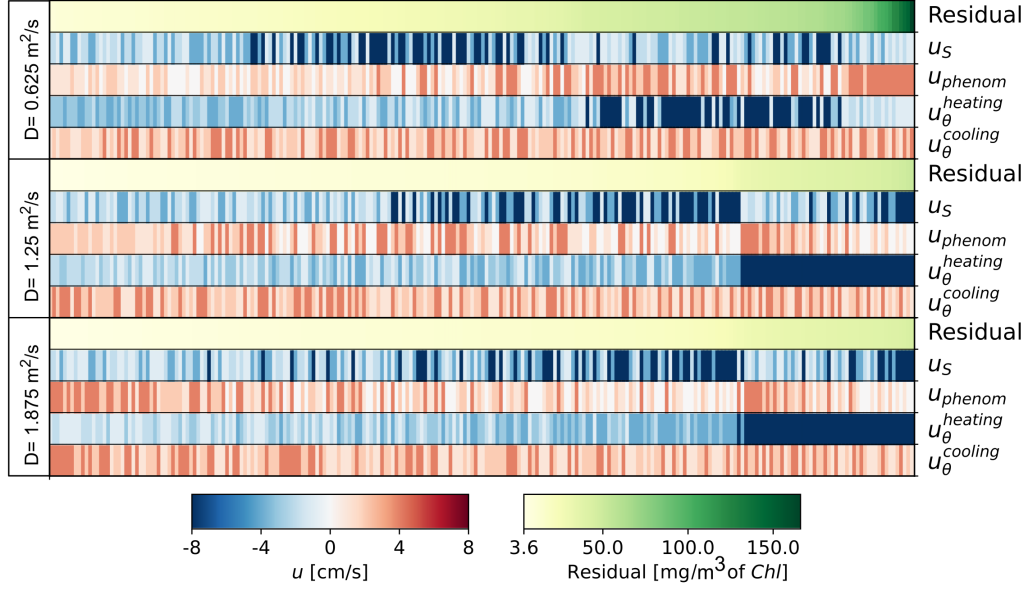


Figure A.14: Residual in  $\text{mg/m}^3$  of Chlorophyll between the biomass cross-shore profiles measured during the survey and the profiles obtained in the model for each tested combination of advective fluxes ( $u_{\theta}^{onshore}$ ,  $u_{\theta}^{offshore}$ ,  $u_s$  and  $u^*$ ). For each model run, the total residual is as an average of the residuals calculated individually across each transect. Model runs are divided per each tested value of diffusion coefficient  $D$ , and ordered from the combination with the lowest residual (left) to the ones with the highest residuals (right). The magnitude of each advective flux is represented by color, red corresponding to onshore fluxes and blue to offshore fluxes.

847 The model performs on average better for the two higher values of cross-shore  
848 diffusivities. The best agreement between modelled and measured biomass  
849 distribution is found for the intermediate value of cross-shore diffusivity  $D$   
850  $= 1.25 \text{ m}^2/\text{s}$ , together with the following magnitudes of advective forcings:  
851  $u_{\theta}^{onshore} = 2 \text{ cm/s}$ ,  $u_{\theta}^{offshore} = -2 \text{ cm/s}$ ,  $u_s = -2 \text{ cm/s}$  and  $u^* = 2 \text{ cm/s}$ . The  
852 modelled cross-shore distribution for this set of parameters is shown in Fig.

853 **A.15.**

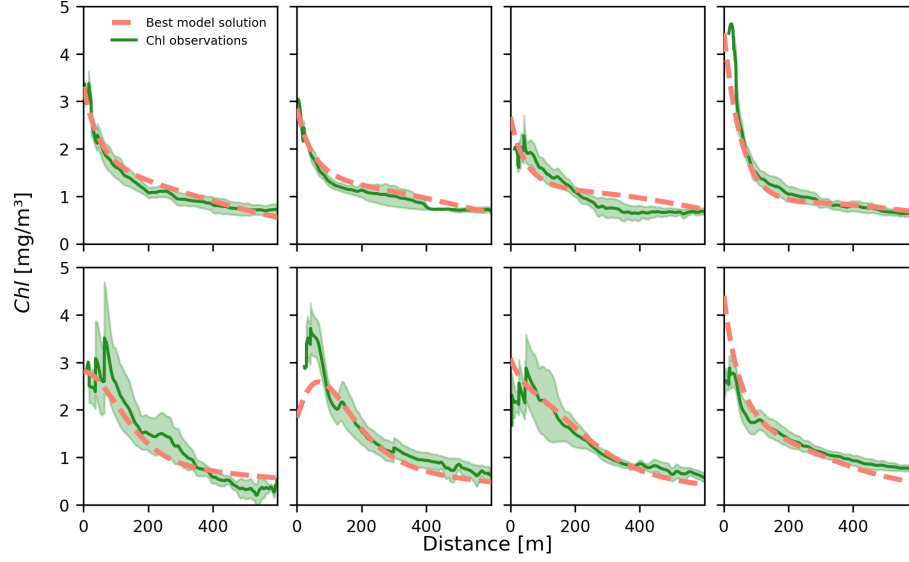


Figure A.15: Comparison between observed and modelled cross-shore biomass distribution for the set of parameters corresponding to the lowest residual ( $D = 1.25 \text{ m}^2/\text{s}$ ,  $|u_{\theta}^{onshore}| = |u_{\theta}^{offshore}| = |u^*| = 2 \text{ cm/s}$ ,  $|u_S| = 2 \text{ cm/s}$ )

854 *Appendix A.11. Alternative scenarios: exploration of the individual role of*  
 855 *the physical forcings*

856 The model predicts that the daily patterns observed in the biomass distribu-  
 857 tion emerge from the dynamical interplay between physical forcings of differ-  
 858 ent origin (wind, temperature and salinity gradients). We can thus investi-  
 859 gate how the increase, the weakening or the disappearance of these forcings  
 860 would affect the biomass pattern. To do so, we considered the set of parame-  
 861 ters that best fit the field data ( $D = 1.25 \text{ m}^2/\text{s}$ ,  $|u_{\theta}^{onshore}| = |u_{\theta}^{offshore}| = |u^*|$   
 862  $= 2 \text{ cm/s}$ ,  $|u_S| = 1 \text{ cm/s}$ ) and modified individually the contribution of each  
 863 forcing, except for the phenomenological current  $u^*$  which is left unchanged.  
 864 We considered the following modifications of the forcings: an increase and

865 decrease by a 50% of the wind strength, and the selective disappearance of  
866 either the temperature or the salinity driven flows.

867 The exploration of these alternative scenarios highlights the critical role of  
868 wind in counteracting biomass dispersion driven by buoyancy flows. A 50%  
869 reduction in wind strength leads to a pronounced afternoon flush of biomass,  
870 effectively halving the total biomass retained within the most nearshore stripe  
871 (Fig.??a-b). Conversely, increasing the wind strength by 50% doubles the  
872 biomass accumulation in the same region. Further offshore, the impact of  
873 wind on total biomass fluctuations diminishes. This is consistent with the  
874 expectation that buoyant flows weaken exponentially with distance from the  
875 coastline, as their driving gradients operate over a scale of only a few hun-  
876 dred meters (Fig.A.13). As a result, beyond this nearshore region, biomass  
877 modulation becomes increasingly independent of buoyancy-driven flows, and  
878 the wind's counterbalancing influence becomes less significant.

879 Overall, our findings demonstrate that wind is the primary driver of the  
880 observed biomass accumulation patterns, with its strength modulating the  
881 amplitude of daily biomass variations. Selectively eliminating one of the two  
882 buoyancy-driven flows (Fig. ??c-d) introduces deviations in the afternoon  
883 biomass variability relative to the observed data. In the case of the temper-  
884 ature driven flow, we can appreciate the role of the differential cooling flow  
885 in maintaining the biomass accumulation in the nearshore zone in the early  
886 morning (7:00 AM, Fig.A.13). Salinity gradients, on the other hand, play a

887 more marginal role in modulating the physico-biological interaction of this  
 888 particular system, where the salinity driven flow kicks in with a few hours  
 889 delay with respect to the one driven by differential heating.

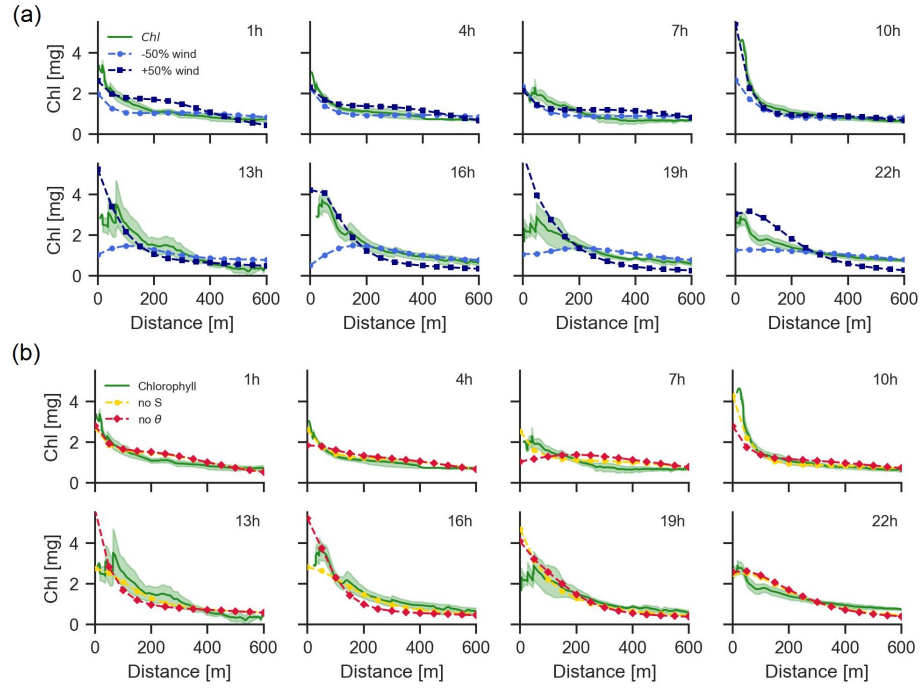


Figure A.16: