

The Cross-Shelf Regime of a Wind-Driven Supercritical River Plume

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7 ABSTRACT: River plumes are a dominant forcing agent in the coastal ocean, transporting tracers
8 and nutrients offshore and interacting with coastal circulation. In this study we characterize the
9 novel ‘cross-shelf’ regime of freshwater river plumes. Rather than remaining coastally-trapped
10 (a well-established regime), a wind-driven cross-shelf plume propagates for tens to over one hun-
11 dred kilometers offshore of the river mouth while remaining coherent. We perform a suite of
12 high-resolution idealized numerical experiments that offer insight into how the cross-shelf regime
13 comes about and the parameter space it occupies. The wind-driven shelf flow comprising the
14 geostrophic along-shelf and the Ekman cross-shelf transport advects the plume momentum and
15 precludes geostrophic adjustment within the plume, leading to continuous generation of internal
16 solitons in the offshore and upstream segment of the plume. The solitons propagate into the
17 plume interior, transporting mass within the plume and suppressing plume widening. We exam-
18 ine an additional ultra-high resolution case that resolves submesoscale dynamics. This case is
19 dynamically consistent with the lower resolution simulations, but additionally captures vigorous
20 inertial-symmetric instability leading to frontal erosion and lateral mixing. We support these
21 findings with observations of the Winyah Bay plume, where the cross-shelf regime is observed
22 under analogous forcing conditions to the model. The study offers an in-depth introduction to the
23 cross-shelf plume regime and a look into the submesoscale mixing phenomena arising in estuarine
24 plumes.

25 SIGNIFICANCE STATEMENT: In this study, we characterize a novel regime of freshwater river
26 plumes. Rather than spreading near to or along the coast, under certain conditions river plumes
27 may propagate away from the coast and remain coherent for tens to over one hundred kilometers
28 offshore. Cross-shelf plumes provide a mechanism by which freshwater and river-borne materials
29 may be transported into the open ocean, especially across wide continental shelves. Such plumes
30 carry nutrients critical for biological productivity offshore and interact with large-scale oceanic
31 features such as the Gulf Stream. We use high-resolution numerical modeling to examine how the
32 cross-shelf regime arises, and support our findings with observational evidence. We also study the
33 mixing phenomena and fluid instabilities evolving within such plumes.

34 1. Introduction

35 A salient feature of the coastal ocean is that cross-shelf gradients of pressure and density are
36 an order of magnitude (or more) higher than the corresponding along-shelf gradients, which,
37 through geostrophic adjustment, results in predominantly along-shelf circulation and transport
38 (Brink 2016). This highly polarized pressure and density distribution is disrupted by localized
39 sources of brackish water running off from estuary or river mouths. The resulting density anomaly
40 frequently reaches $O(10 \text{ kg m}^{-3})$ (Yankovsky et al. 2022), a buoyancy forcing magnitude seldom
41 seen in the open ocean. The buoyant outflow tends to spread radially in the form of a gravity
42 current from its source (at least when unopposed by the ambient shelf circulation, e.g., O'Donnell
43 (1990)), forming a nearly circular coastal buoyant plume (Garvine 1984). The complex dynamics
44 associated with the advancing front of the buoyant plume are reviewed by O'Donnell (2010),
45 while an example of modern observations of frontal processes can be found in Delatolas et al.
46 (2023). The periphery of the plume reaches a geostrophic, or sometimes a gradient wind balance
47 (Yankovsky and Chapman 1997), resulting in an anticyclonic flow pattern commonly referred to
48 as an anticyclonic bulge.

49 There is a tendency for the anticyclonic bulge to grow indefinitely (Nof and Pichevin 2001).
50 However, this tendency breaks down due to the development of various instabilities (e.g., Oey and
51 Mellor (1993); Jia and Yankovsky (2012); Izett and Fennel (2018)). The excessive buoyant water is
52 transported downstream so that the anticyclonic bulge approaches a spatial limit, albeit fluctuating
53 with time. Sharples et al. (2017) related the spatial scale of the anticyclonic bulge to the shelf

width as a metric for the delivery of riverine nutrients across the shelf break, a process of primary importance for deep-ocean ecosystems.

The buoyant flow ultimately exits the anticyclonic bulge and evolves into a coastally-trapped buoyant current propagating alongshore in the direction of Kelvin wave phase (hereinafter referred to as downstream) and exhibiting a geostrophic balance in the cross-shelf direction. This natural downstream propagation can be reversed under the influence of wind forcing or deep ocean circulation impinging on the shelf. The wind-induced reversal of the buoyant coastal current is associated with upwelling conditions, such that the buoyant water propagates not only upstream, but also offshore, in the surface boundary layer. The offshore transport is sustained by Ekman dynamics. Regardless of the direction (downstream vs. upstream), both patterns typically reveal significant alongshore extension of buoyant water which retains its contact with the coastline (e.g., Hickey et al. (2009)).

This study addresses a less frequent regime, where the plume detaches from the coast near the mouth and crosses the shelf at an oblique angle as an elongated tongue of buoyant water. This regime was described and referred to as a cross-shelf plume in the recent observational study of the Winyah Bay plume off the South Carolina coast, USA, by Yankovsky et al. (2022), where similar plumes at other locations were also briefly discussed. The important finding of Yankovsky et al. (2022) is that cross-shelf plumes are formed under the influence of light to moderate upwelling-favorable wind, and the buoyant layer remains supercritical in terms of the internal Froude number tens of km away from the source (an estuary mouth). Froude number is defined as:

$$F = U_s / C_i, \quad (1)$$

where U_s is near-surface velocity magnitude typically averaged over the buoyant layer depth, and C_i is internal gravity wave phase speed. The supercriticality is maintained by superposition of the inherent flow field associated with the plume density anomaly (which tends to achieve a geostrophic balance) and the wind-induced Ekman drift. Yankovsky et al. (2022) assessed supercriticality by solving an internal wave (IW) eigenvalue problem and recovering the gravest IW mode dispersion curve for an arbitrary buoyancy frequency profile and a vertically sheared mean current. This is a more accurate approach than a slab-like representation of a buoyant layer typically utilized for

81 internal Froude number estimates (e.g., Branch et al. (2020), Geyer and Ralston (2011), Hetland
82 (2010), MacDonald and Geyer (2005), among many others).

83 Supercritical buoyant outflows are frequent in nature, and can be identified by a liftoff, or
84 detachment from the bottom, of a buoyant layer near the mouth, where an internal hydraulic jump
85 occurs. Such outflows are also characterized by high Burger number (Bu) at the mouth, where Bu
86 is the ratio of the baroclinic Rossby radius to the mouth width:

$$\text{Bu} = \sqrt{g'h}/(fW). \quad (2)$$

87 Here $g' = |g\Delta\rho|/\rho_0$ is the reduced gravity, where g is gravitational acceleration, $\Delta\rho$ is the buoyant
88 layer density anomaly relative to the ambient seawater of density ρ_0 , h is the buoyant layer thickness,
89 f is the Coriolis parameter, and W is the mouth width. The significance of the supercritical regime
90 in the plume near field (as defined by Garvine (1984); see also Horner-Devine et al. (2015)) was
91 discussed mostly in terms of the buoyant layer mixing and entrainment (Hetland 2005, 2010). In
92 fact, the internal Froude number can be related to the Richardson number (O'Donnell 1990), with
93 the latter governing Kelvin-Helmholtz instability, and consequently the mixing and entrainment
94 across the interface of the buoyant layer. The role of supercritical flow in setting the plume's spatial
95 structure has seldom been addressed. The notable exceptions are a series of papers by Garvine
96 (1984, 1987) and O'Donnell (1990), as well as some references therein. In these studies, the plume
97 structure is determined by propagation properties of internal waves (expressed mathematically as
98 characteristic lines in a corresponding solution) which lead to the formation of trailing fronts and
99 interior jumps within the plume, also referred to as interior fronts. Under unforced conditions, the
100 plume cannot remain supercritical over long distances from the mouth due to rapid momentum
101 dissipation, but superimposed light wind stress can change this situation (Yankovsky et al. 2022).

102 In this study, we address the formation of the cross-shelf plume by means of idealized numerical
103 modeling. As demonstrated by Yankovsky et al. (2022), the cross-shelf plume does not exhibit
104 substantial transverse (lateral) spreading which implies some inherent dynamics sustaining the
105 coherent plume structure. A set of numerical experiments retaining the essential ingredients of a
106 cross-shelf plume is intended to delineate these dynamics. The rest of the paper is organized as
107 follows: Section 2 describes the model configuration and governing parameters. Section 3 presents
108 the model results beginning with a coastally-trapped plume and then examining the emergence,

109 properties, and parameter space of the cross-shelf plume regime. Section 4 presents an ultra-high
110 resolution simulation and examines submesoscale mixing dynamics. Section 5 presents satellite
111 observations supporting the model findings, and addresses some inevitable simplifications of the
112 model. Section 6 concludes the paper.

113 2. Numerical model

114 We use the non-hydrostatic MITgcm (Marshall et al. 1997) to solve the incompressible Boussinesq
115 equations in z -coordinates. The model domain is 3-dimensional and lies on an f -plane, with
116 Coriolis parameter $f = 8 \cdot 10^{-5} \text{ s}^{-1}$. A non-linear equation of state is used to solve for density based
117 on McDougall et al. (2003). A Laplacian viscosity with constant horizontal and vertical coefficients
118 $A_h = 10 \text{ m}^2 \text{ s}^{-1}$ and $A_z = 10^{-4} \text{ m}^2 \text{ s}^{-1}$, respectively, ensures numerical stability by dissipating energy
119 near the grid scale such that the grid Reynolds number is near order one. Temperature and salinity
120 diffusivities are both set to $K_h = 10^{-6} \text{ m}^2 \text{ s}^{-1}$ in the horizontal and $K_z = 10^{-6} \text{ m}^2 \text{ s}^{-1}$ in the vertical.
121 The advection scheme is third-order direct-space-time with a flux limiter. Buoyant outflows with
122 sharp density gradients undergo significant numerical diffusion even when the spatial resolution is
123 relatively high (e.g. Ralston et al. (2017)). For this reason, we apply low diffusivity coefficients in
124 order to prevent excessive erosion of the pycnocline and frontal interfaces, and thus to permit the
125 development of instabilities and internal waves.

126 The zonal and meridional extents of the domain are 60 and 100 km, respectively, and maximum
127 depth is 25 meters. Horizontal grid spacing is $dx = 100 \text{ m}$ and vertical grid spacing is $dz = 0.1 \text{ m}$.
128 A higher-resolution run with $dx = 15 \text{ m}$ and lower horizontal viscosity ($A_h = 1.5 \text{ m}^2 \text{ s}^{-1}$) is also
129 performed as a sensitivity test and to examine smaller-scale dynamical influences. The domain is
130 a periodic meridional channel with zonal open boundary conditions discussed below. The origin
131 of the coordinate system is in the southwestern corner with x , y , and z axes pointing eastward,
132 northward, and upward, respectively, and corresponding velocity components (u, v, w) . There is a
133 10 km-wide strip of land along the western boundary, with a coastal wall located at $x = 10 \text{ km}$. The
134 land is cut through by a zonal inlet of width W , which is set to 1 km in the standard configuration
135 ($W = 5, 10 \text{ km}$ are also tested). Buoyant inflow enters the domain through this inlet. The water
136 depth at the coastal wall and within the inlet is 5 m; it linearly increases to 10 m at $x = 12 \text{ km}$ and
137 then to its maximum value of 25 m at $x = 37 \text{ km}$. In one case the alongshore-uniform coastline

is replaced with a coastal promontory (hereinafter, referred to as a cape) similar to the coastline geometry of Winyah Bay. In this case, the coastline linearly recedes to $x = 0$ km over the 15-km alongshore distance both north and south of the mouth.

The model is forced by a buoyant inflow Q specified at the head of the inlet ($x = 0$ km) with depth-independent inflow velocity. The zonal boundary conditions are such that inflow from the western boundary is balanced by outflow from the eastern boundary. Aside from the inflow, there is no normal flow on the western boundary. The coastal wall uses free slip conditions (zero stress). There is a free surface and the bottom stress is parameterized via quadratic bottom drag with a coefficient of 0.005. The buoyant inflow mimics the buoyancy forcing from Winyah Bay as reported by Yankovsky and Voulgaris (2019) and Yankovsky et al. (2022) and corresponds to the time-averaged riverine freshwater discharge $Q_r = 800 \text{ m}^3 \text{ s}^{-1}$. The buoyant inflow is brackish, with the depth-averaged inflow salinity $\bar{s}_i = 18$; s_i increases linearly from 12 at the surface to 24 at the bottom at $x = 0$. The ocean ambient salinity is $s_0 = 34$. The inflow oscillates with a period of 12.4 h (an M2 tidal harmonic). Thus, Q at the inlet head is defined as:

$$Q = Q_r \frac{s_0}{s_0 - \bar{s}_i} [1 + \sin(\omega t)], \quad (3)$$

where t is time and ω is the tidal frequency. During the first tidal cycle, no salinity anomaly is prescribed at the inlet head, that is, inflow has the ambient ocean salinity. A constant in time meridional wind stress of 0.03 Pa is applied in the standard model configuration (higher values are also explored). This wind stress direction generates an eastward Ekman transport and leads to upwelling conditions at the western (coastal) boundary.

A total of 11 model runs are reported here (with more sensitivity tests completed) and their configurations are summarized in Table 1. The nominal duration of each model run is 3.1 days (74.4 h), which is 6 full tidal cycle, 5 with buoyancy forcing. However, the model blew up earlier, at ~ 2.3 days in model run SW10 (the strongest wind with standard A_z), most likely due to “drying” of the uppermost grid cell. Several diagnostics are used to characterize the model results. The vertical profile of salinity s is reduced to the equivalent freshwater layer thickness:

$$h_f = \int_{-D}^0 \frac{s_0 - s}{s_0} dz, \quad (4)$$

TABLE 1. Summary of model experiments, with bold text indicating the unique feature of each experiment. The standard case (S) is listed first. Other abbreviations are: NW = no wind, NT = no tides, M5 = 5 km mouth, M10 = 10 km mouth, W5 = 0.05 Pa winds, W10 = 0.1 Pa winds, V1 and V2 are alternate viscosity values, HR = high resolution, C = cape geometry. R_f is defined in Eq. 18 as the fraction of the freshwater volume advected past the channel midpoint at $x = 35$ km. Values of R_f are given at two times: (2.3, 3.1) days.

Model Run	Tidal Discharge	W [km]	Coastal Cape	τ [Pa]	A_z m ² s ⁻¹	Δx [m]	Bu	R_f
S	Yes	1	No	0.03	10^{-4}	100	7.31	(0.19, 0.45)
SNW	Yes	1	No	0	10^{-4}	100	7.09	(0.0, 0.0)
SNT	No	1	No	0.03	10^{-4}	100	7.57	(0.18, 0.44)
SM5	Yes	5	No	0.03	10^{-4}	100	1.12	(0.02, 0.29)
SM10	Yes	10	No	0.03	10^{-4}	100	0.51	(0.01, 0.25)
SW5	Yes	1	No	0.05	10^{-4}	100	7.51	(0.45, 0.61)
SW10	Yes	1	No	0.1	10^{-4}	100	7.91	(0.60, N/A)
V1W10	Yes	1	No	0.1	$2 \cdot 10^{-4}$	100	8.10	(0.21, 0.37)
V2W10	Yes	1	No	0.1	10^{-3}	100	8.74	(0.00, 0.02)
SHR	Yes	1	No	0.03	10^{-4}	15	7.87	(N/A, N/A)
SC	Yes	1	Yes	0.03	10^{-4}	100	7.55	(0.13, 0.42)

Since the plume occupies only a small fraction of the total water depth D , the internal gravity wave phase speed C_i is deduced from h_f as $C_i = \sqrt{g'h_f}$, where g' corresponds to the freshwater density anomaly. Burger number (Eq. 2) at the mouth is estimated as $Bu = C_i/(fW)$, with h_f (which varies significantly across the mouth for model runs SM5 and SM10) taking its maximal value across the mouth and averaged over the last tidal cycle. Since there is no stratification in the ambient ocean, C_i becomes zero there and the internal Froude number becomes infinite. For this reason, we utilize the inverse Froude number:

$$F_i = C_i/U_s. \quad (5)$$

U_s is averaged over the buoyant layer of thickness h defined following Arneborg et al. (2007) as:

$$h = \frac{2 \int_{-D}^0 (\rho_0 - \rho) z dz}{\int_{-D}^0 (\rho_0 - \rho) dz}. \quad (6)$$

176 Note $F_i < 1$ corresponds to the supercritical regime (F_i should only be considered within the plume
177 as the ambient ocean is unstratified).

178 **3. Results**

179 Here we present the suite of model simulations summarized in Table 1 aimed at studying the
180 emergence and breakdown of the cross-shelf plume regime. We begin by considering a case
181 without wind forcing. After establishing the dynamical regime of the unforced plume, we add a
182 light upwelling-favorable wind forcing to our model. This leads to the striking emergence of the
183 cross-shelf plume regime, which persists under a variety of inflow conditions. Finally, we consider
184 under which forcing conditions the cross-shelf regime breaks down. To further support these
185 findings we perform an ultra-high resolution simulation of the nominal cross-shelf plume case. We
186 find that although the dynamics broadly converge between the lower and higher resolution cases,
187 the latter offers novel insights into the submesoscale mixing phenomena leading to mixing within
188 the plume. We devote the last subsection of the results to examining the higher-resolution case and
189 the submesoscale mixing dynamics captured there.

190 *a. The Unforced, Coastally-Trapped Plume*

194 We start by considering a rotational plume under no wind forcing (case SNW in Table 1).
195 The buoyant inflow conditions at the mouth can be characterized as supercritical and high Bu ,
196 implying strong momentum advection, separation of the buoyant layer from the frictional bottom,
197 and initial radial spreading of buoyant water from the mouth. Such buoyancy forcing represents
198 a classical subject in plume research and has been extensively studied (e.g., Chao and Boicourt
199 (1986); Garvine (1987); Oey and Mellor (1993)). Rotational, buoyant plumes under no wind
200 forcing consist of four regions (see Horner-Devine et al. (2015), Figure 2 for a schematic). The
201 source region extends from the river mouth to the liftoff location, where the buoyant plume loses
202 contact with the bottom. The jet-like near-field region follows; here the momentum of the plume
203 dominates over other forces, resulting in strong mixing. In the mid-field region, the plume is
204 increasingly dominated by Earth's rotation, forming an anticyclonic circulation (bulge). Finally,
205 the far-field region of the plume is a geostrophic coastal current that carries the riverine water along
206 the coast in the direction of Kelvin wave propagation.

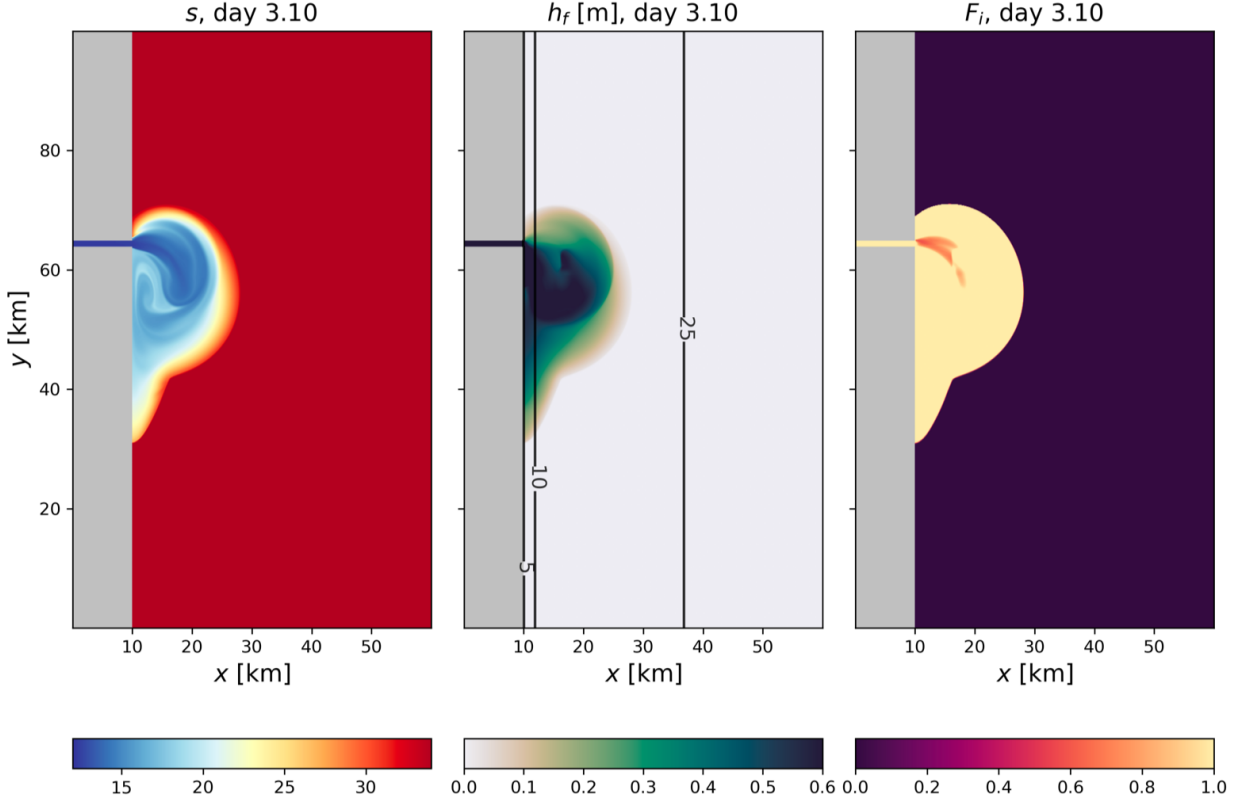


FIG. 1. No wind forcing case (SNW in Table 1), left to right: surface salinity, equivalent freshwater layer thickness (h_f), and inverse Froude number (F_i , with $F_i < 1$ being supercritical). Isobaths are shown in the middle panel.

The plume shown in Figure 1 behaves as we expect from a high Bu, supercritical inflow, initially forming a thin layer that spreads radially with the ebbing tidal pulse (seen in the salinity field). The plume's front advances following each tidal cycle in the form of an internal bore. The plume is shaped by Earth's rotation, developing a mid-field anticyclonic bulge and a far-field coastal current after just five tidal cycles. The flow is supercritical only near the mouth but then transitions to the subcritical regime ($F_i > 1$). The plume spreads as a thin, 2 m deep buoyant layer (Figure 2), which is roughly comparable to the Ekman depth

$$h_E = \sqrt{(2A_z/f)}. \quad (7)$$

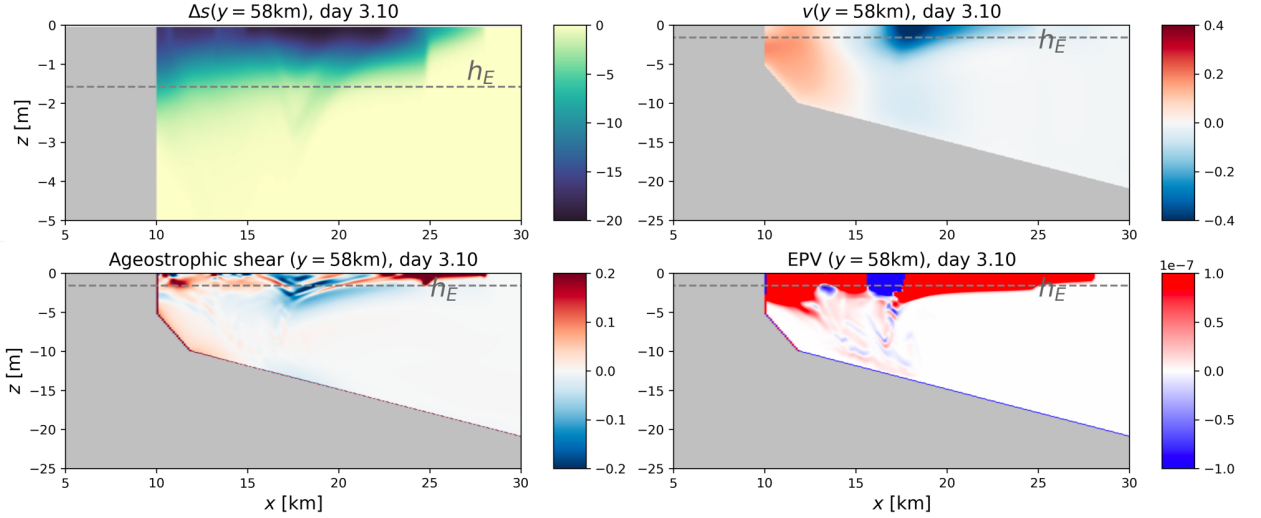


FIG. 2. No wind forcing case (SNW), vertical transects are taken across the plume at $y = 58$ km and 3.10 days. Upper left panel is the salinity anomaly Δs with Ekman depth h_E marked, upper right is the alongshore velocity component v , lower left is the ageostrophic shear, and lower right is the Ertel potential vorticity. Note the smaller vertical scale of the upper left plot.

Having a sufficiently small h_E in our model ensures that the wind forcing imposed in subsequent numerical experiments does not artificially deepen the surface boundary layer to significantly exceed the buoyant layer depth. The alongshore velocity v is characterized by strong vertical shear. However, the shear deviates from the geostrophic shear component; this is expected in rotational plumes, but at a later stage of their development and farther away from the source. Ageostrophic shear is the difference between vertical shear ($\partial v / \partial z$) and thermal wind shear ($\partial v_g / \partial z$):

$$\frac{\partial v}{\partial z} - \frac{\partial v_g}{\partial z} = \frac{\partial v}{\partial z} - \frac{-g}{f\rho_0} \frac{\partial \rho}{\partial x}. \quad (8)$$

The ageostrophic shear field (Figure 2, lower left) shows the presence of beam-like structures of alternating sign. These beams are reminiscent of inertial-symmetric instability (ISI), which develops in frontal regions with strong geostrophic shear (Grisouard 2018). Pure symmetric instability (SI) extracts energy from the flow through the vertical geostrophic shear production (GSP) term

$$\text{GSP} = -\overline{\hat{v}\hat{w}} \frac{\partial \overline{v_g}}{\partial z}, \quad (9)$$

where an overline indicates a spatial average over the SI scale and hats are deviations from the average, whereas pure inertial instability (InI) extracts energy through the lateral shear production (LSP) term (Thomas et al. 2013; Wenegrat and Thomas 2020):

$$\text{LSP} = -\hat{v}\hat{u}\frac{\partial\overline{v_g}}{\partial x}. \quad (10)$$

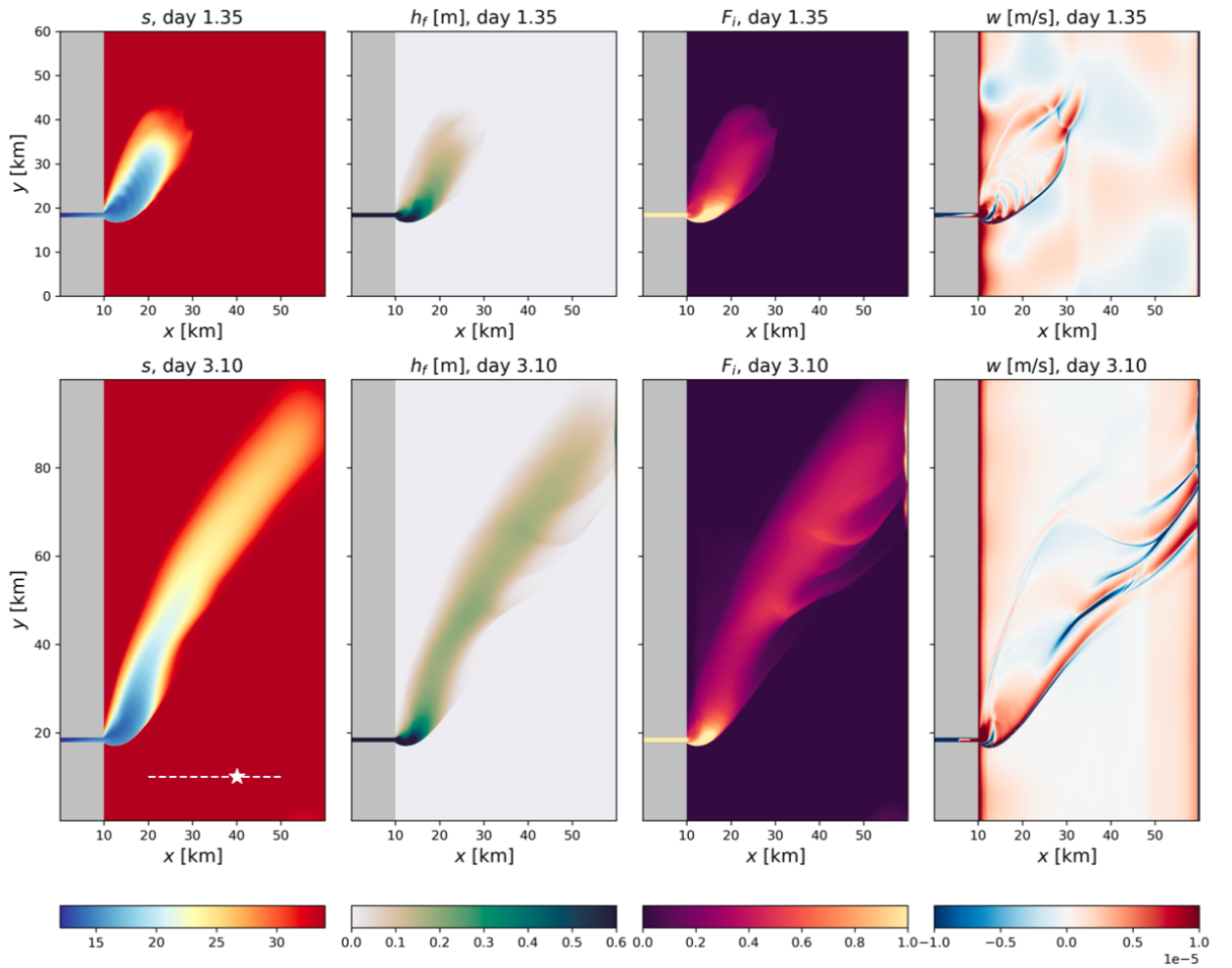
SI, InI, and ISI hybrids (energized through a combination of GSP and LSP) may be diagnosed by computing the Ertel potential vorticity (EPV):

$$\text{EPV} = (f\hat{\mathbf{k}} + \nabla \wedge \mathbf{u}) \cdot \nabla b, \quad (11)$$

where buoyancy is defined as $b = -g\rho/\rho_0$.

When the EPV is opposite in sign to the Coriolis parameter and the flow is gravitationally stable, the condition for instability is met (Hoskins 1974). In Figure 2 we see that the beam-like structures originate in the area of negative EPV, indicating ISI. Based on Equations 9-11, we identify SI, InI, and hybrid ISI depending on the orientation of the front relative to the background geostrophic current (discussed in greater detail in Section 4). Interestingly, the salinity anomaly (Figure 2, upper left) shows beams of freshwater propagating downwards and being mixed. ISI provides a mixing mechanism that erodes and deepens the plume front. Although SI was previously identified in realistic regional simulations (Ayouche et al. 2020, 2021) and in idealized process studies (Lv et al. 2020), it arose in the presence of wind forcing and with several times coarser spatial resolutions. Ayouche et al. (2020, 2021) found that the Hoskins instability criterion for SI is satisfied in frontal regions within the Gironde plume in the Bay of Biscay. However, their model had horizontal resolutions of 200-400 meters and vertical resolutions on the order of 10s of meters. Similarly, horizontal resolutions in Lv et al. (2020) were 1-2 km. As a result, the vertical shears and velocity beams driven by SI (and ISI hybrids) evolving on horizontal scales of hundreds of meters and vertical scales of 1-10s of meters were under-resolved (Bachman et al. 2017). In our domain, we achieve resolutions of 0.1 meters vertically and 100 m horizontally (15 meters in our ultra-high resolution case, Section 4). We are thus able to capture the vertical ISI beams as well as some of the secondary mixing and frontal erosion. Thus, we have considered a case with no wind forcing to simulate the well-studied coastally-trapped regime ubiquitous among natural river plumes. We

255 established that our model accurately represents the dynamics arising in a rotational river plume.
 256 Additionally, owing to the high resolution of our model we identified a new mixing pathway arising
 257 in this regime. The buoyant inflow advects negative EPV into the domain, which gives rise to
 258 inertial-symmetric instability. ISI creates diagonal ageostrophic velocity beams that lead to mixing
 259 of plume tracers and deepening of the plume front. We will next consider what happens when a
 260 light northward (upwelling-favorable wind) is incorporated into our model.



261 FIG. 3. Standard case without tides (SNT), fields are shown at days 1.35 (upper row) and 3.10 (lower row).
 262 Left to right: surface salinity, equivalent freshwater layer thickness (h_f), inverse Froude number ($F_i < 1$ is
 263 supercritical), and vertical velocity (w). The star and dashed line in the lower left panel indicate where the
 264 Ekman velocity and geostrophic velocity are computed in Figure 4.

b. The Cross-Shelf Regime

1) OVERVIEW

After establishing the coastally-trapped plume regime in the prior section, we add a temporally constant wind forcing of 0.03 Pa to our model (case SNT, Table 1). The wind is directed northward, thus upwelling-favorable, and discharge is constant in time following the initial ramp-up (no tidal modulations). The evolution of the plume is shown in Figure 3. We see a stark contrast with the alongshore regime discussed prior. Rather than forming a bulge and deflecting back towards the coast in a geostrophic current, the plume detaches from the coast and spreads upstream and offshore, curving slightly to the right with distance from the mouth. It retains a tight transverse structure (seen in the salinity field), that is, it does not broaden substantially with distance from its source. Salinity increases towards the flanks of the plume due to mixing, but we do not see the radial spreading characteristic of the prior SNW case.

We observe radiation of solitary internal waves (Apel et al. 2007) at the upstream (northward) flank of the plume and propagating in the downstream (southward) direction. These solitons are seen in the freshwater content field as curving lines of higher h_f values. Due to superposition of the wind-induced drift, we also see that the plume remains supercritical as it moves offshore ($Fi < 1$, Figure 3). By transporting mass and momentum into the interior and downstream part of the plume, the solitons suppress mixing at the northward edge and prevent the plume from spreading laterally. The plume remains coherent and supercritical over the cross-shore extension of the numerical domain through this mechanism, described in greater detail below. Figure 3 shows two time snapshots: day 1.35, when the first soliton is clearly formed, and day 3.1 (the end of integration). Solitons can also be identified in the vertical velocity component w as localized areas of high negative values (shown in blue in Figure 3), residing within the plume.

2) BACKGROUND FLOW DYNAMICS

We now address the dynamics of the cross-shelf plume in detail, first considering the background flow onto which the plume is superimposed. We choose locations outside of the plume, shown as a transect and star in Figure 3. The background flow dynamics include: (i) surface wind-induced eastward Ekman flow; (ii) an along-channel geostrophic current driven by the cross-channel pressure gradient (which results from the Ekman transport divergence/convergence at the

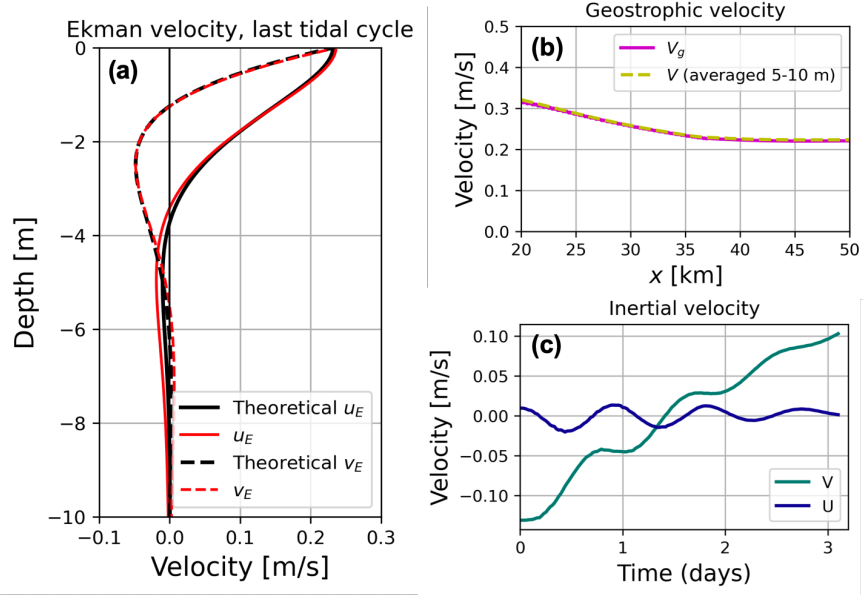


FIG. 4. Standard case without tides (SNT): (a) theoretical and observed Ekman velocity components (u_E, v_E) as a function of depth and averaged over the last tidal cycle at the starred location in Figure 3, (b) theoretical geostrophic velocity v_g and observed v (alongshore velocity averaged over 5-10 meters) also averaged over the last tidal cycle, and (c) inertial velocity components at the same location as (a) at 12.5 m depth, shown as a function of time.

western/eastern wall, respectively), and (iii) decaying near-inertial oscillations. These features are shown in Figure 4. In panel (a), we computed the observed Ekman velocity components by averaging the horizontal velocities over the last tidal cycle and subtracting the mid-depth velocity. The vertical structure of these velocities is in near-perfect agreement with the well-known Ekman solution in the surface boundary layer outside of the plume:

$$(u_E, v_E) = \frac{\tau e^{z/h_E}}{\rho_0 \sqrt{A_z f}} (\cos(\pi/4 + z/h_E), \sin(\pi/4 + z/h_E)). \quad (12)$$

The location is shown as a star in Figure 3 and is not affected by the presence of buoyant water (and hence the associated baroclinic velocity shear). The geostrophic along-channel current also generates compensating westward Ekman transport near the bottom (not shown). The along-channel velocity averaged over 5-10 meters depth, outside of the surface and bottom boundary layers and away from coastal walls, is shown in panel (b) to be in geostrophic balance with the

309 barotropic cross-shore pressure gradient set by the sloping sea surface. Lastly, in panel (c) we
 310 obtained the mid-depth velocity as a function of time and subtracted its temporal mean, for the
 311 same location as panel (a). Both u and v velocity components exhibit weak velocity oscillations
 312 with periods close to the inertial period of 21.82 h. These features validate the physical behavior
 313 of our model, as it exhibits all the theoretical expectations for flow dynamics outside the plume.

314 3) MOMENTUM BALANCE WITHIN THE PLUME AND INTERNAL WAVE GENERATION

315 The momentum balance within the plume region is assessed at 1.35 days, the time when soliton
 316 radiation begins (Figure 5). To eliminate the surface wind stress term, we consider the first grid cell
 317 below the surface (0.1 m depth). The leading-order terms both outside and inside the plume are the
 318 cross-shelf pressure gradient, Coriolis force, and vertical divergence of shear stresses. Although the
 319 viscous term combines both horizontal and vertical components, the former is negligible except for
 320 the upwind frontal interface (not shown). The inertial term reveals internal wave motions: radiating
 321 from the mouth and radiating southward from the upstream (northward) flank of the plume. In
 322 addition, there are streaks of momentum advection along the upwind part of the plume, negative
 323 in x -momentum balance and positive in y -momentum balance, which overlap with the maximal
 324 downward vertical velocity associated with the soliton generation (second column, Figure 5).

336 The momentum balance shown in Figure 5 is too complex to explain the process of internal wave
 337 generation in the offshore part of the plume, although it suggests that the momentum advection
 338 terms may be responsible. We further investigate this process through several steps. In Figure
 339 6a, we show the near surface flow field (u', v') where the wind-driven ambient flow (u_a, v_a) is
 340 subtracted:

$$(u', v') = (u - u_a, v - v_a). \quad (13)$$

341 Here $(u_a(y), v_a(y))$ are taken outside of the plume and the flow field is again considered at 0.1
 342 m depth (first interior grid point). Primed variables represent plume dynamics and along with the
 343 freshwater layer topography show the disturbance at $\sim(20 < x < 30 \text{ km}, 35 < y < 45 \text{ km})$ which
 344 subsequently propagates as an internal soliton. In Figure 6b, the same field is shown for a model
 345 run where the momentum advection is turned off. Without the nonlinear terms, no perturbations
 346 occur at the advancing edge of the plume. Horizontal current vectors at the advancing edge are
 347 outward of the plume (near perpendicular to local h_f contours), consistent with the dynamics of the

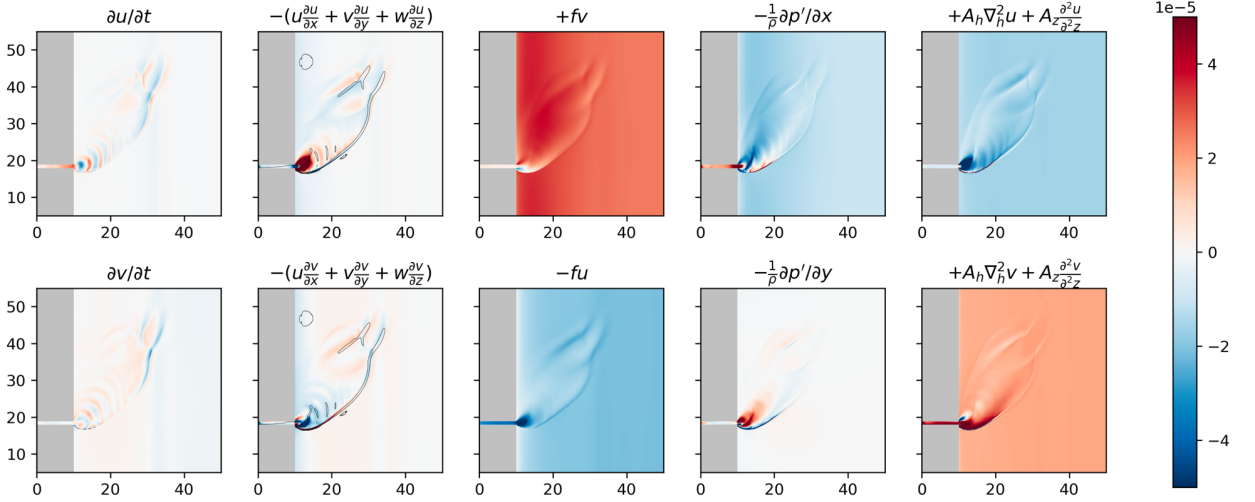


FIG. 5. Standard case without tides (SNT): top row shows the zonal momentum budget and bottom row shows the meridional momentum budget at 1.35 days at 0.1 m depth. Terms are denoted in each subplot's title, left to right: time tendency, advection, Coriolis, pressure gradient, and viscous terms. The second column includes an isoline of vertical velocity w (at the value $-2 \cdot 10^{-6} \text{ m s}^{-1}$).

gravity current, while behind the leading edge (closer to the mouth) velocity adjusts geostrophically and becomes more aligned with local h_f contours. It is clear at this point that the internal wave radiation in the offshore segment of the plume is due to the momentum advection associated with the ambient wind-driven flow field, since neither the linear solution of this model experiment, nor the nonlinear unforced plume (Figure 1) give rise to internal waves offshore.

Next, we construct a simplified momentum balance with nonlinear terms which is applied to the linear solution's flow field along the transect $x = 20.5 \text{ km}$, $35 < y < 45 \text{ km}$ shown in Figure 6b. This transect approximately represents the locus of the maximum freshwater content, the most downwind-protruding part of the plume, and is also the area where the internal soliton originates in the nonlinear solution (Figure 6a). We assume that the wind-induced ambient flow is independent of the alongshore coordinate y . Furthermore, at the offshore location of this transect, the Ekman dynamics are fully developed, and u_a is assumed to be spatially uniform. Subtracting equations for u_a and v_a from the total momentum balance equations yields:

$$\frac{\partial u'}{\partial t} + (u' + u_a) \frac{\partial u'}{\partial x} + (v' + v_a) \frac{\partial u'}{\partial y} - f v' = -\frac{1}{\rho} \frac{\partial p'}{\partial x} + A_z \frac{\partial^2 u'}{\partial z^2}, \quad (14)$$

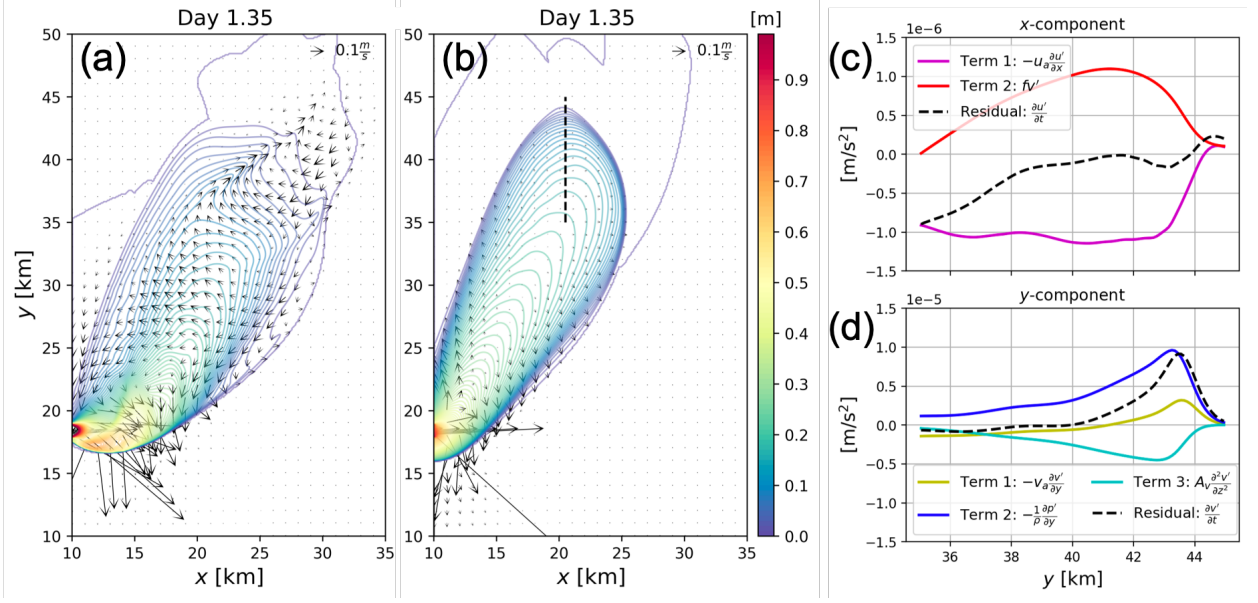


FIG. 6. (a) Standard case without tides (SNT): Freshwater layer thickness h_f (shown as colored contours) and the near-surface velocity field (shown as vectors) associated with the plume at 1.35 days when solitons develop on both sides of the divergent segment of the plume. The velocity associated with the plume is computed by subtracting the surface velocity defined along a zonal transect at $y = 60$ km (outside the plume) from each zonal transect in the domain. (b) Same as (a), but for a simulation where the momentum advection in the model has been turned off. Dashed black line shows the transect for which the momentum balance in panels (c)-(d) is computed. (c) Momentum balance based on Equation 16. (d) Momentum balance based on Equation 17.

$$\frac{\partial v'}{\partial t} + (u' + u_a) \frac{\partial v'}{\partial x} + u' \frac{\partial v_a}{\partial x} + (v' + v_a) \frac{\partial v'}{\partial y} + f u' = -\frac{1}{\rho} \frac{\partial p'}{\partial y} + A_z \frac{\partial^2 v'}{\partial z^2}. \quad (15)$$

Here p is the pressure; the horizontal viscosity is ignored because it is important only at the upwind front, and the vertical momentum advection is ignored because the ambient flow has small vertical velocities at this offshore location. Next, we assume that since velocities at the leading edge of the plume are smaller than the wind induced advection, the nonlinear terms associated with the plume dynamics alone (products of primed variables) can also be discarded. Lastly, along the locus of the h_f maximum several other simplifications can be made: $\frac{\partial p'}{\partial x}$, u' , $\frac{\partial u'}{\partial y}$, $\frac{\partial v'}{\partial x} = 0$. These assumptions for velocity and its derivatives are based on the near-normal velocity orientation with respect to h_f contours at the leading edge of the plume. With these assumptions, the momentum balance equations are reduced to:

$$\frac{\partial u'}{\partial t} = -u_a \frac{\partial u'}{\partial x} + f v', \quad (16)$$

$$\frac{\partial v'}{\partial t} = -v_a \frac{\partial v'}{\partial y} - \frac{1}{\rho} \frac{\partial p'}{\partial y} + A_z \frac{\partial^2 v'}{\partial z^2}. \quad (17)$$

370 We treat the linear solution (Figure 6b) as the mean, unperturbed state of the plume, and we
 371 estimate nonlinear terms in Equations 16 and 17 from this state. If these nonlinear terms are
 372 permitted, they should lead to the perturbation similar to that shown in Figure 6a. Without the
 373 advective terms, the momentum balance will lead to geostrophic adjustment (which has already
 374 formed closer to the mouth in Figure 6b): the outward v momentum will be turned offshore through
 375 the Coriolis effect ($\frac{\partial u'}{\partial t} > 0$ in the x -momentum balance), and the northward pressure gradient force
 376 in the y -momentum balance will be balanced by the resulting Coriolis force $f u'$. The balance will
 377 not be purely geostrophic due to the contribution of the vertical eddy viscosity.

378 This situation changes drastically when the nonlinear terms are present (Figure 6c, d). In the
 379 x -momentum balance, the offshore advection (through the Ekman transport) of u' counteracts
 380 the Coriolis force, and hence prevents geostrophic adjustment (that is, clockwise turning of the
 381 velocity vector is suppressed). In fact, for $y < 40$ km, the acceleration is negative, giving rise to
 382 the along-front flow in the onshore direction, opposite to geostrophic flow. In the y -momentum
 383 balance, the advective term changes sign and becomes positive (adding to the northward pressure
 384 gradient force) at $y = 41$ km, and its magnitude exceeds the vertical viscosity term at the edge of
 385 the plume. As a result, the v' acceleration matches (or even exceeds) the pressure gradient force
 386 for $y > 43$ km. These two conspicuous features are seen in the offshore and downwind segment
 387 of the plume ($x > 20$ km, $35 < y < 45$ km) in the nonlinear solution (Figure 6a): the onshore flow
 388 along h_f contours (against the geostrophic direction) and the northward/offshore jet crossing the
 389 front and exiting the plume. This flow pattern results in a crest-like structure of the buoyant layer
 390 thickness, which subsequently radiates southward as an internal soliton.

391 Finally, it should be noted that the internal wave generation is expected to be highly sensitive to
 392 the vertical eddy viscosity, since this is the only term opposing the combined effect of the pressure
 393 gradient and momentum advection in Equation 17 at the edge of the plume. Its role will be further
 394 discussed in the next section.

Thus, we summarize the maintenance of the cross-shelf plume regime as follows:

1. The advection of plume momentum by the ambient wind driven flow prevents geostrophic adjustment after the passage of the plume's leading edge. Instead, the offshore part of the plume continuously radiates internal solitons (see animations in supplementary material).
2. Solitons radiate into the plume (upwind), transporting mass and momentum and preventing downwind diffusion of the plume. Solitons are suppressed beyond the plume's boundaries as they cannot propagate into unstratified water. Thus, the plume remains supercritical and retains its coherence as it propagates further offshore.

4) PARAMETER SPACE OF THE CROSS-SHELF REGIME

Here we perform a sensitivity study to delineate the parameter space under which the cross-shelf regime persists. We add a tidally modulated estuarine inflow to our model configuration (Figures 7 and 8). We investigate whether variable discharge, which enhances nonlinear advection of momentum in the near field, has an effect on the dynamics previously identified for the constant discharge case. We then modify several model parameters to assess the conditions under which the cross-shelf regime persists vs. breaks down. Case S in panel (a) is the standard case and is analogous to the previously considered SNT (see Table 1). The plume structure is more complex than in Figure 3 and comprises tidal sub-plumes partially separated by gaps in salinity and h_f . These gaps gradually merge in the older part of the plume. The number of solitons seen in the h_f panel (Figure 8a) has increased compared to the case with steady-state inflow (Figure 3); in this case each tidal cycle produces a soliton. In the two next cases (Figures 7-8, b-c) the mouth width increases such that Bu drops to 1.12 for $W = 5$ km and to 0.51 for $W = 10$ km. When $Bu < 1$ in an unforced (no wind) plume, there is a tendency for geostrophic adjustment of the inflow, with the nonlinear momentum advection becoming negligible. In panels (b-c) we see that tidal sub-plumes become progressively less distinct with increasing W . However, the cross-shelf regime still persists even when $Bu < 1$, albeit with slightly reduced overall offshore extension of the buoyant layer. The mechanism described above for the maintenance of the cross-shelf regime by soliton radiation continues to operate and the resulting plume retains its narrow transverse structure. The cross-shelf plume regime breaks down in the case with stronger wind stress and enhanced vertical eddy viscosity/diffusivity coefficients (Figures 7-8d, case V2W10 in Table 1). In this case the

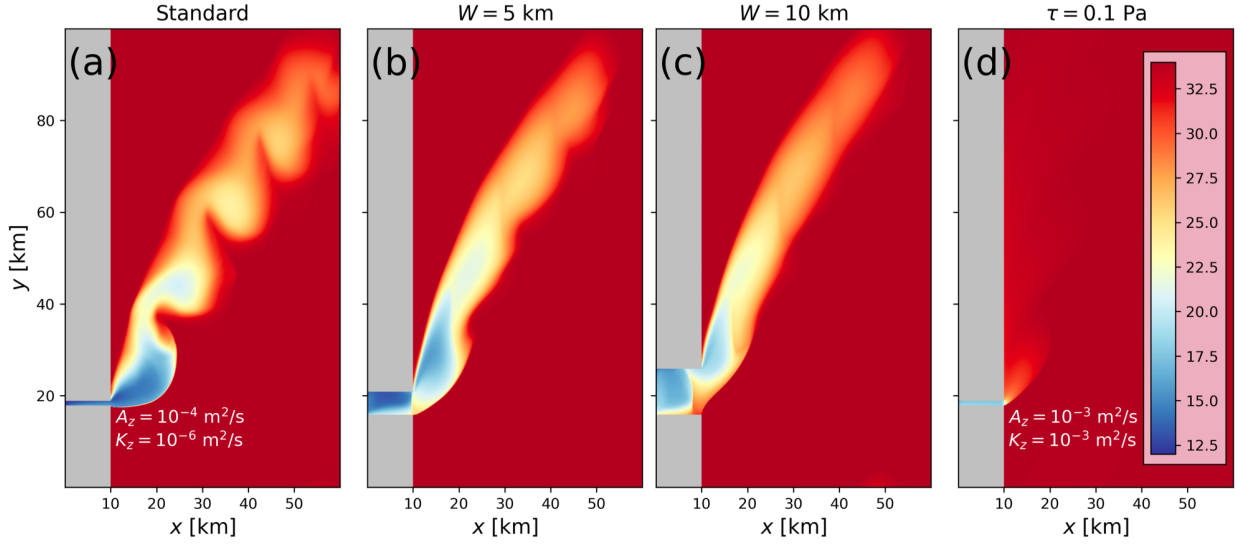


FIG. 7. Surface salinity at 3.1 days for four cases: (a) standard (S), (b) 5 km wide mouth (SM5), (c) 10 km wide mouth (SM10), (d) highest wind forcing and higher viscosity and diffusivity values (V2W10). In cases S, SM5, and SM10 the vertical viscosity is $A_z = 10^{-4} \text{ m}^2 \text{ s}^{-1}$ and vertical diffusivity is $K_z = 10^{-6} \text{ m}^2 \text{ s}^{-1}$, while in case V2W10, $A_z = 10^{-3} \text{ m}^2 \text{ s}^{-1}$ and $K_z = 10^{-3} \text{ m}^2 \text{ s}^{-1}$.

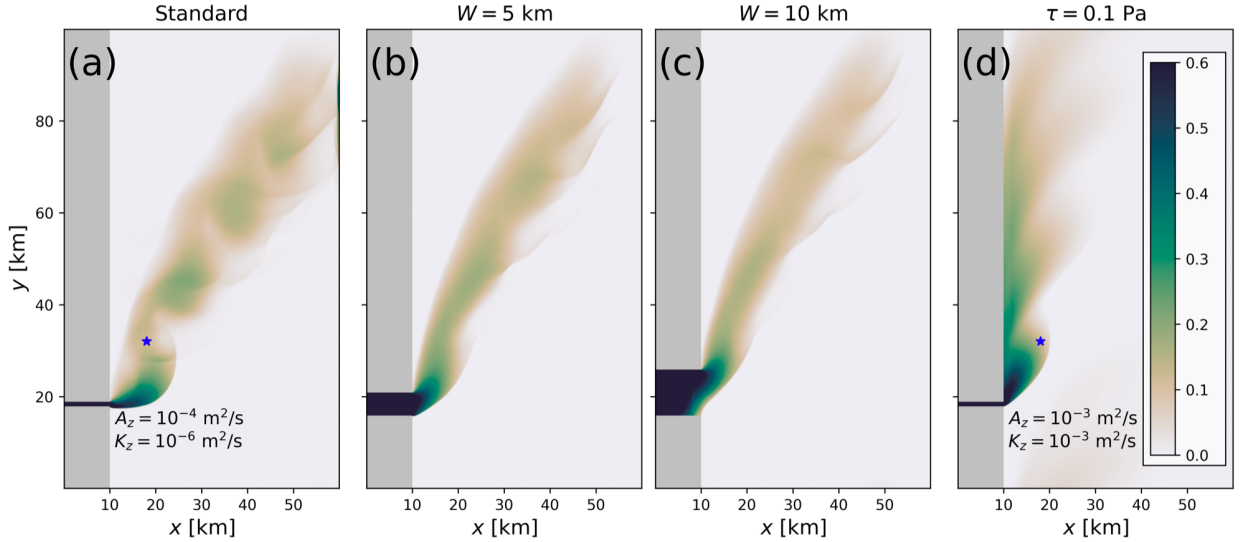


FIG. 8. Equivalent freshwater layer thickness h_f at 3.1 days for the same cases as Figure 7.

plume spreads predominantly alongshore in the upstream direction, with some offshore Ekman transport. The salinity signal is very weak (the same color scheme is used for all panels in Figure 7), and the largest concentration of freshwater is along the coast.

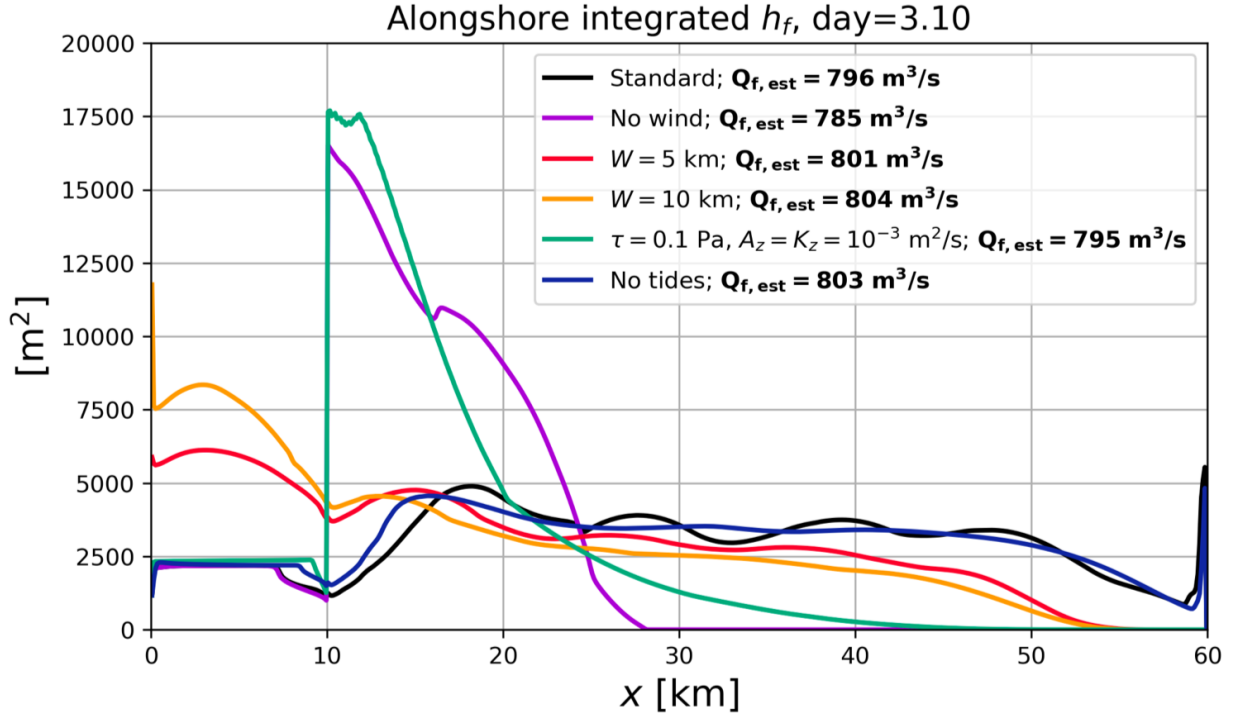


FIG. 9. Alongshore integrated equivalent freshwater layer thickness h_f as a function of offshore distance x for a suite of cases, top to bottom: S, SNW, SM5, SM10, V2W10, and SNT. Shown in the legend is the apparent freshwater discharge for each case (this quantity should be $800 \text{ m}^3 \text{ s}^{-1}$, and is used as a measure of model performance).

Cross-shore transport of freshwater associated with the cross-shelf plume is delineated in Figure 9, where h_f (Equation 4) is integrated along the channel. This integral, when divided by the duration of the inflow (five tidal cycles) also yields an offline estimated freshwater discharge ($Q_{f,est}$), which can be compared against the nominal freshwater discharge of $800 \text{ m}^3 \text{ s}^{-1}$ to gauge model error. In all but one case shown in Figure 8 the discrepancy is less than 1%, which is a very commendable performance of the model. The only case with larger than 1% discrepancy is the SNW case without wind stress (785 vs. nominal $800 \text{ m}^3 \text{ s}^{-1}$), where the horizontal advection of salinity is the lowest among all cases considered.

The freshwater distribution across the channel with and without tidally modulated inflow is nearly identical (Figure 9), with some freshwater accumulating at the eastern wall of the channel. The tendency for offshore freshwater accumulation increases with wind stress increase, for a given eddy viscosity value (cases SW5 and SW10, see below). The cross-shore freshwater delivery is reduced

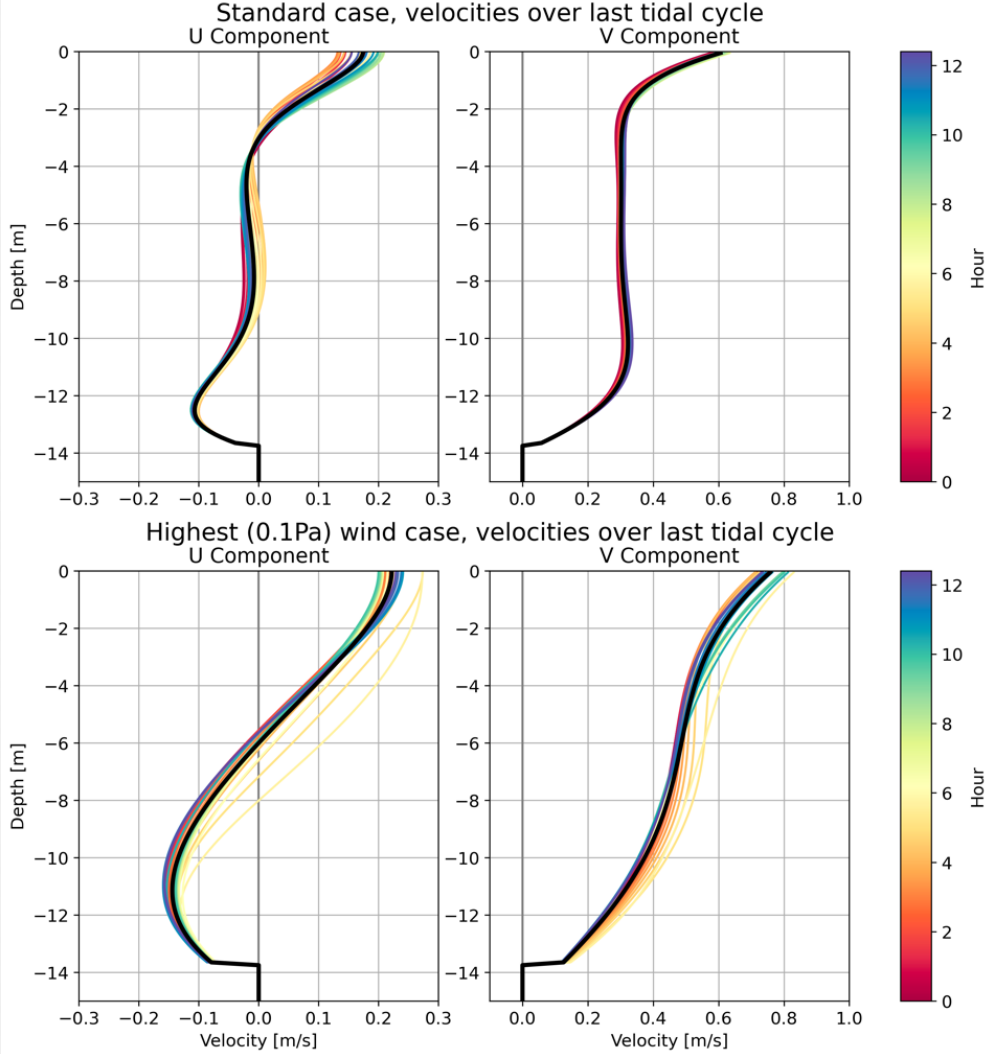


FIG. 10. Vertical profiles of horizontal velocity components (u, v) within the plume at the location marked as a blue star in Figure 8. Velocity profiles are shown as snapshots every half hour for the last tidal cycle (color corresponds to time of the snapshot); the time-averaged profile is shown in black. The top row is for case S (standard) and the lower row for case V2W10 (highest wind forcing and higher viscosity and diffusivity values).

in the cases with wider mouths, but this reduction is primarily due to accumulation of freshwater in the estuary. To better quantify the cross-shore transport of buoyant water, the following coefficient for the offshore freshwater delivery is used:

$$R_f = \frac{\int_{x=35}^{x=60} h_f dx}{\int_{x=10}^{x=60} h_f dx}, \quad (18)$$

454 where R_f is the ratio of the freshwater volume advected past the channel midpoint at $x = 35$ km to
 455 the total amount of freshwater advected past $x = 10$ km, which is the mouth location. See Table 1 for
 456 values of R_f for each experiment. We find R_f weakly depends on the estuarine Bu (i.e., decreases
 457 from 0.29 to 0.25 as Bu decreases from 1.12 to 0.51, cases SM5 and SM10) and the presence of
 458 tidal modulation (i.e., increases from 0.44 to 0.45 between cases SNT and S). On the other hand, R_f
 459 drastically increases with increasing wind stress when A_z and K_z are kept constant, as in cases S,
 460 SW5, and SW10 at 2.3 days (increasing from 0.19 to 0.45 to 0.60 between the cases). The latter is
 461 a somewhat unrealistic proposition as both A_z and K_z are likely to increase with wind stress; these
 462 experiments are simply meant to emphasize the role of the advection when the vertical viscosity and
 463 diffusivity are both low. Once A_z and K_z are increased, R_f is dramatically reduced. In particular,
 464 cases S and V2W10 have almost the same depth-averaged Ekman velocities: 0.23 and 0.24 m
 465 s^{-1} , respectively, which are estimated as U_E/h_E , where $U_E = \tau/(\rho_0 f)$ is the Ekman transport.
 466 However, in the latter case R_f is only 0.02 by the end of integration (very little freshwater crosses
 467 the channel midpoint), while in the S case R_f is 0.45. As discussed, case V2W10 demonstrates
 468 the breakdown of the cross-shelf plume regime.

472 The vertical structure of horizontal currents near the coast in water depths of less than 14 m
 473 is examined for two cases, S and V2W10, in Figure 10. In the standard case S the surface and
 474 bottom boundary layers are well separated by mid-depth flow with no vertical shear, while in the
 475 case with higher A_z , the boundary layers overlap and the flow is continuously sheared from the
 476 surface to the bottom. This regime is often referred to as the inner shelf dynamics (Lentz 2001),
 477 and if A_z further increases, the veering of the horizontal velocity vector with depth can be reduced
 478 and the cross-shelf flow pattern (offshore near-surface and onshore near-bottom) can be severely
 479 weakened. In addition, we see an increased alongshore v -component from the S to the V2W10 case
 480 due to the wind stress increase, while the cross-shelf velocity remains approximately the same due
 481 to deepening of the Ekman layer. Hence, the advection of buoyant water shifts to a predominantly
 482 alongshore pathway and the cross-shelf regime is suppressed.

483 The previously considered cases S, SM5 and SM10 all have the same wind-driven circulation
 484 (and hence, ambient advection), and the plumes are driven by the same freshwater discharge,
 485 making the conditions for internal wave radiation roughly comparable. In case V2W10 the internal
 486 wave activity is completely shut down due to the increased viscosity A_z (Figures 7-8). We now

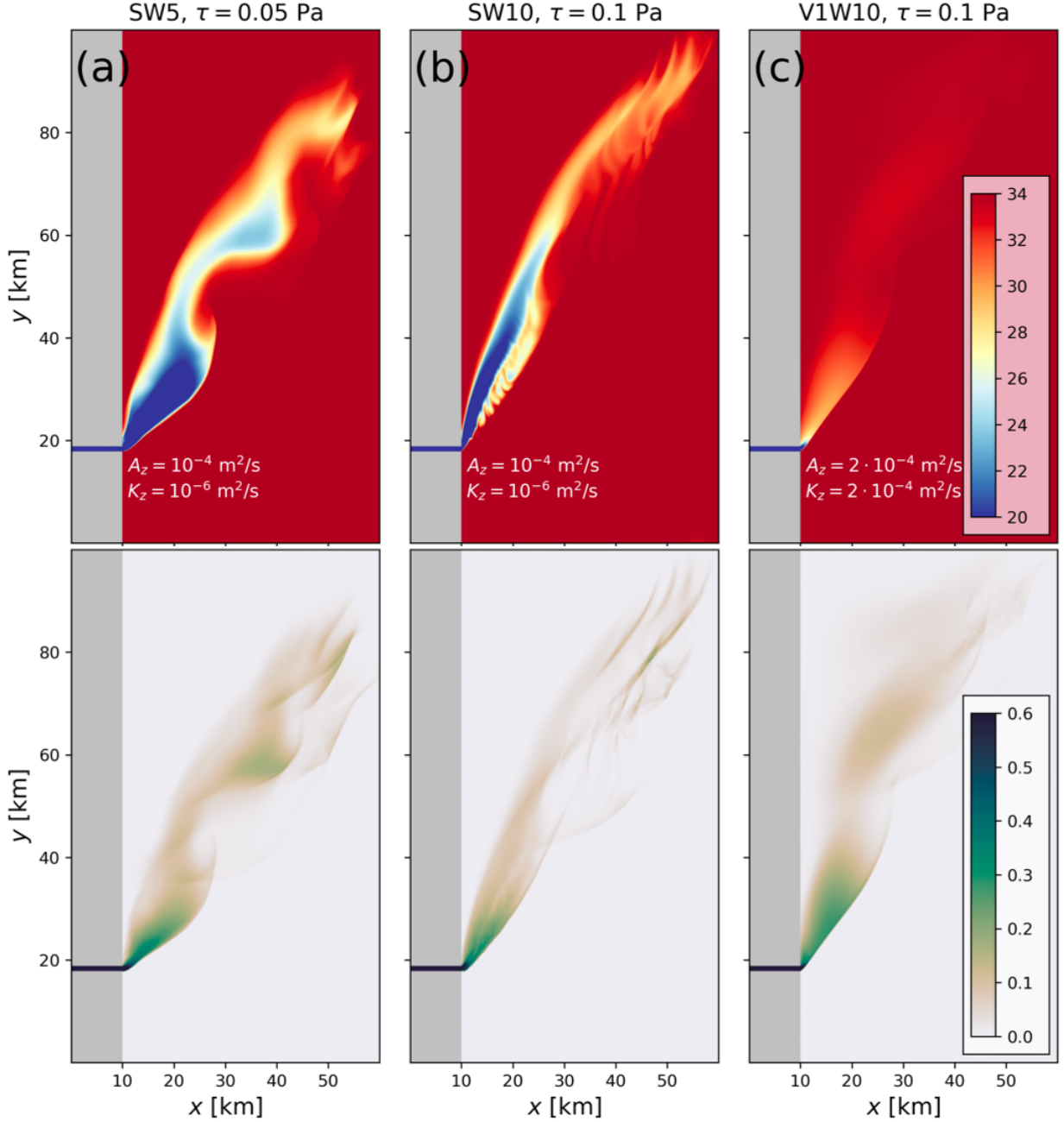


FIG. 11. Surface salinity (top) and equivalent freshwater layer thickness h_f (bottom). The cases are (left to right): SW5 (with 0.05 Pa wind stress), SW10 (with 0.1 Pa wind stress), and V1W10 (with 0.1 Pa wind stress and higher viscosity and diffusivity values, indicated in the figure).

consider cases where both ambient advection and vertical eddy viscosity change (Figure 11) which should significantly affect the internal wave dynamics as discussed in subsection 3. In model runs

SW5 and SW10, the wind stress increases to 0.05 and 0.1 Pa (compared to 0.03 Pa in case S), leading to faster advection both along- and off-shore. This translates into enhanced generation of internal solitons now seen not only in the h_f distribution, but in the surface salinity field as well. In particular for case SW10, most of the freshwater is contained in internal solitons, and the offshore plume dynamics ($x > 30$ km) are now highly nonstationary and nonlinear, without a well defined “bulge”. On the other hand, when both vertical viscosity and diffusivity are moderately increased (case V1W10), internal wave activity is significantly reduced (but still present), while the downwind (northern) edge of the plume becomes more diffuse offshore (seen in the h_f field for $x > 25$ km). These results indicate that the cross-shelf freshwater transport is most sensitive to the magnitude of the zonal wind stress and vertical mixing coefficients.

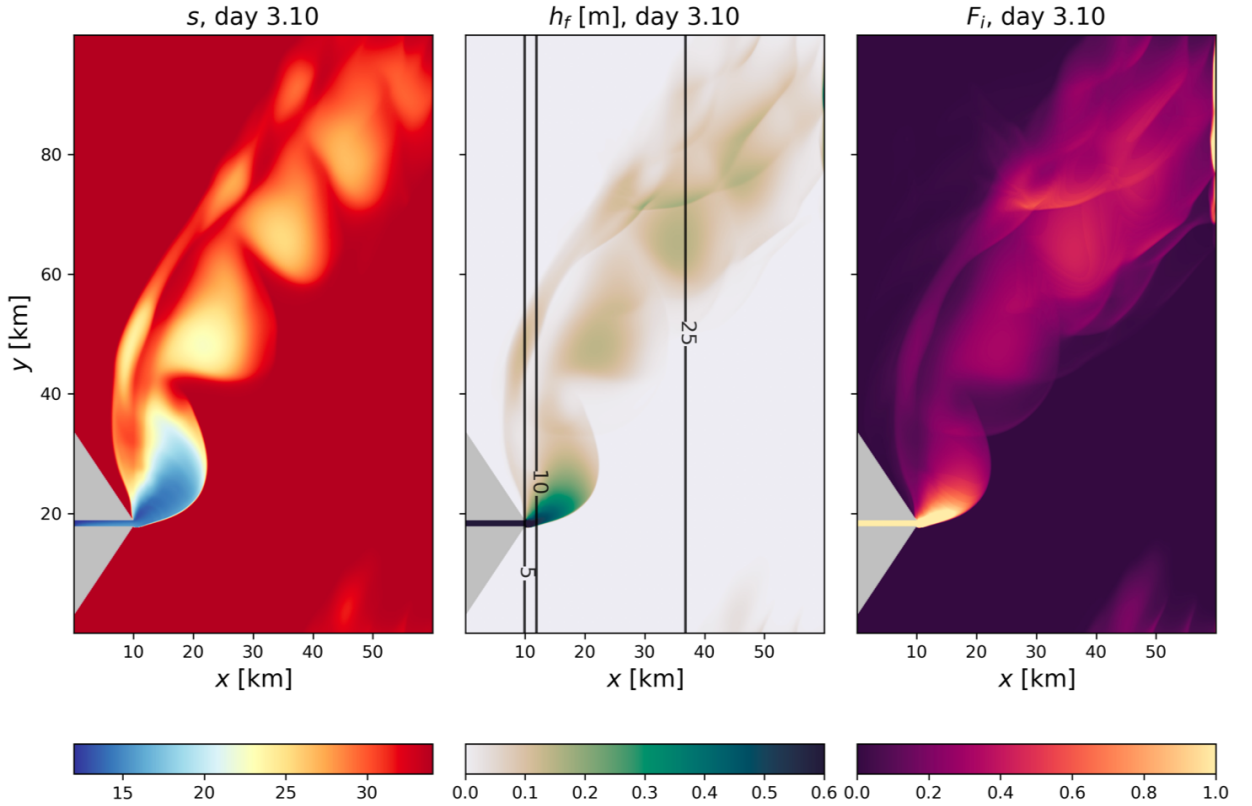


FIG. 12. Case SC, with a coastal promontory corresponding to the idealized geometry of Winyah Bay. Left to right: surface salinity, equivalent freshwater layer thickness (h_f), and inverse Froude number ($F_i < 1$ is supercritical) at 3.10 days.

502 5) THE CASE OF A COASTAL PROMONTORY

508 The modeling study presented here has been motivated by observations of the Winyah Bay Plume.
509 The channel of Winyah Bay runs off into the ocean through a coastal promontory formed by barrier
510 islands on both sides of the channel (not a straight coastline as in the idealized cases discussed
511 thus far). In the model run SC, a similar coastal promontory is introduced such that the coastline
512 outside this feature lies at $x = 0$ km (Figure 12). Otherwise, the model configuration remains the
513 same as in the standard case S. Numerous prior studies have demonstrated that a cape-like coastal
514 feature can lead to a separation of the coastally-trapped buoyant current from the coastline (e.g.,
515 Igeta et al. (2017); Whitney (2023); Zhou and Wu (2023)). The resulting plume structure is more
516 complex than in previous cases but there is no additional offshore deflection of the plume compared
517 to case S and the overall offshore delivery of buoyant water in cases S and SC is similar (in fact R_f
518 is slightly higher in S, see Table 1). The plume now has a longitudinal gap (or discontinuity) in the
519 h_f distribution with two asymmetric parts: the main body and the thin upstream part (the latter
520 is formed when the discharge decreases through the tidal cycle). Both parts of the plume radiate
521 internal solitons in the upwind direction, so that the overall pattern is rather complex in the older
522 parts of the plume farther offshore. We conclude that the existence of a coastal promontory is not
523 crucial for establishing the cross-shelf plume regime of the Winyah Bay outflow.

524 4. An Ultra-High Resolution Case

525 We now present results of an ultra-high resolution simulation, SHR, identical to case S but with
526 a horizontal resolution an order of magnitude finer (Table 1). The vertical resolution is kept the
527 same (0.1 m), but the horizontal resolution is now 15 m (compared to 100 m in case S) in the x
528 and y directions. The simulation has a time step of 1 second and is computationally expensive; we
529 therefore only extend the integration to 2.07 days (three tidal cycles). This simulation is performed
530 for two reasons. The first is to verify convergence with case S, to serve as further evidence
531 (in addition to the observations presented in the next section), that the dynamics captured by our
532 idealized model suite are robust. The second is to uncover additional insights into the submesoscale
533 mixing phenomena arising in the cross-shelf plume regime.

536 Figure 13 shows surface fields of salinity, freshwater content h_f and inverse Froude number
537 (as in Figures 3 and 12). We observe the same mechanisms maintaining the cross-shelf plume

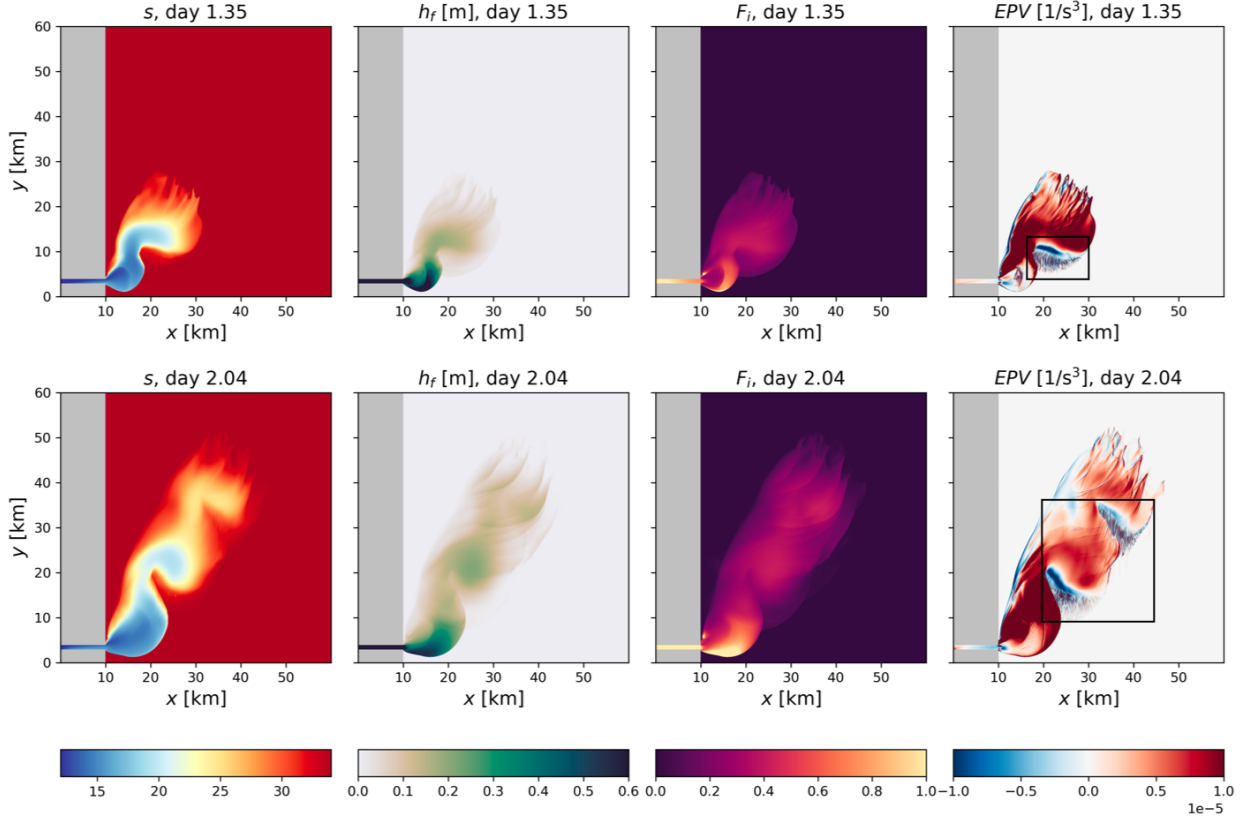


FIG. 13. Case SHR (high-resolution case); note that the domain has been slightly decreased in size to accommodate the higher computational cost. Top row shows fields at 1.35 days, lower row shows fields at 2.07 days (end of simulation). Left to right: surface salinity, equivalent freshwater layer thickness (h_f), inverse Froude number ($F_i < 1$ is supercritical), and Ertel potential vorticity (EPV, with negative values indicating unstable regions). Black rectangles indicate regions where closeup views will be shown in Figure 14.

regime – solitons radiate from the northern part of the plume and propagate upwind, maintaining its coherence. Here, the solitons are better resolved and disintegrate to form trains of high-frequency internal waves (although we are still unable to resolve the breaking of the internal waves). The distribution and transport of h_f is very similar to cases S and SNT; compared to Figures 12 and 3 the SHR case even appears to have a higher freshwater layer thickness moving offshore. We also see that the plume maintains supercriticality as in the previously discussed cases. All of these features indicate that our results do converge across resolutions, further validating our model choices.

We now examine the Ertel potential vorticity (Equation 11, Figure 13); as in Section 3a, negative values indicate regions unstable to submesoscale inertial, symmetric, or hybrid inertial-symmetric

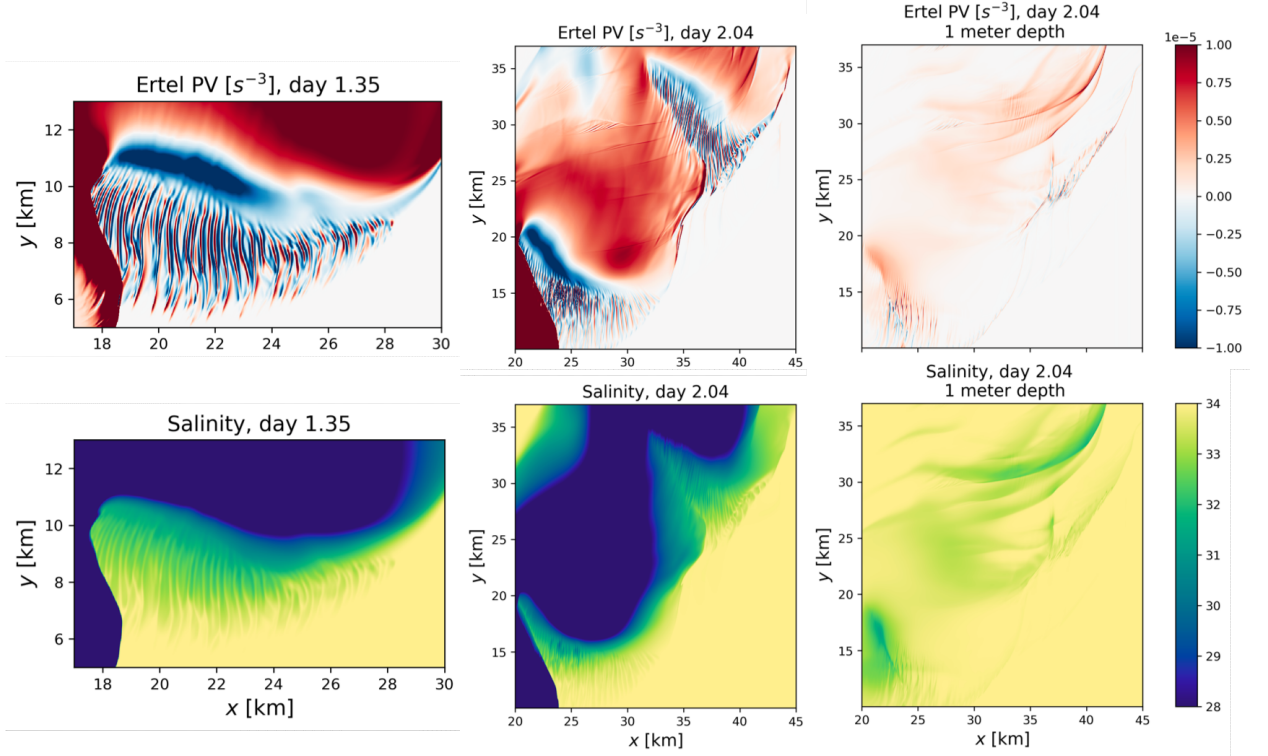


FIG. 14. Closeup views from Figure 13; top row shows Ertel potential vorticity (EPV) and lower row shows salinity. Fields are shown at 1.35 and 2.07 days. The rightmost plots show EPV and salinity at a 1 m depth.

instability (InI, SI, ISI). We observe regions of negative EPV between tidal sub-plumes, along with beam-like structures emanating for several km from the negative EPV regions. As the tidal subplumes propagate into the domain, their boundaries create a sharp density front with the ambient water. Combined with the background geostrophic shear vigorous ISI develops, initiating secondary shear instabilities and mixing. We see contributions both from the lateral and vertical geostrophic shear production terms (Equations 9-10), depending on the evolving orientation of the plume relative to the background geostrophic current. Around 1.35 days, the density front is aligned mostly along the x axis, parallel to $\partial v_g / \partial x$ and a main energy source is the vertical GSP term (Equation 9). The instability has an SI component at this time (likely coexisting with InI/ISI), with particle displacements mostly parallel to isopycnals. Similar SI dynamics were studied by Yankovsky and Legg (2019) but for the case of a dense shelf plume. We note that in the present scenario the instability dynamics are more complex than much of the theoretical literature on SI due to the background flow field, soliton presence, and rapid advection of the plume front offshore.

Furthermore, the front has a curved structure (rather than two-dimensionally symmetric) and its orientation relative to the background sheared flow changes in space and time. As the plume propagates offshore and the density front changes orientation, the LSP term becomes a dominant energy source, corresponding to hybrid ISI. We show a closeup view in Figure 14; the salinity front is undergoing lateral mixing by the ISI confined to the surface; at a depth of 1 meter we only see a small signature of the ISI-driven mixing and the freshwater penetration is dominated by soliton and internal wave activity. This contrasts with case SNW (Figure 2), where the front is more stationary and thus the instability is able to locally propagate deeper. We emphasize that in this complex scenario, distinguishing SI from hybrid ISI is challenging, and we thus simply conclude that there is a vigorous field of submesoscale instability stemming from the combination of velocity shear (geostrophic and perhaps ageostrophic) and density fronts. Our SHR case is able to resolve frontal ISI motions as well as some of the secondary mixing processes, offering insights into the role of ISI in frontal mixing of a plume.

5. Discussion

In this study, we have addressed the formation of a cross-shelf buoyant plume by means of idealized numerical experiments. The model reproduces important features of such a plume: the elongated shape of the buoyant layer which detaches from the coast and crosses the shelf curving gradually offshore with distance from the mouth. An important element for the cross-shelf plume is the detachment of the thin buoyant layer from the coast; in this way the discharged water escapes the nearshore regime of overlapping boundary layers and can be advected offshore by the Ekman transport. This can be achieved under light-to-moderate wind conditions when the wind-induced mixing cannot efficiently erode the stratification of the buoyant layer. Once the wind stress increases, surface and bottom boundary layers merge while the rate of alongshore advection increases such that the buoyant water remains trapped nearshore and propagates upstream (i.e., downwind). Even stronger wind allows the offshore spreading of the buoyant water by Ekman dynamics, but in a more conventional, diffuse manner when the along- and cross-shelf scales of the plume become comparable.

To further illustrate these two distinct regimes, we present two examples of the Winyah Bay plume. One case, observed on April 26, 2019, represents a typical manifestation of the cross-

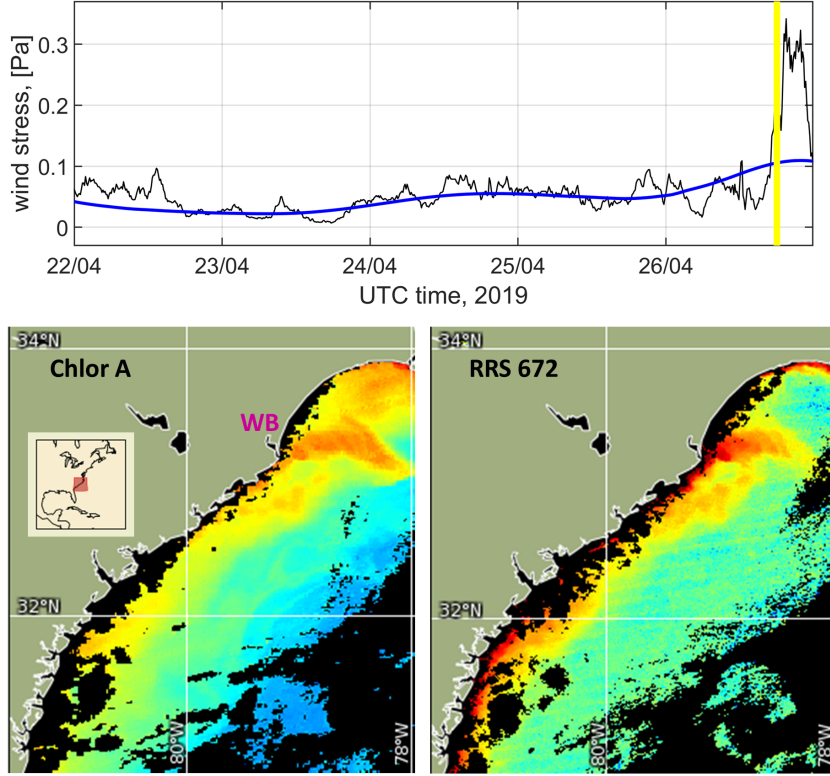
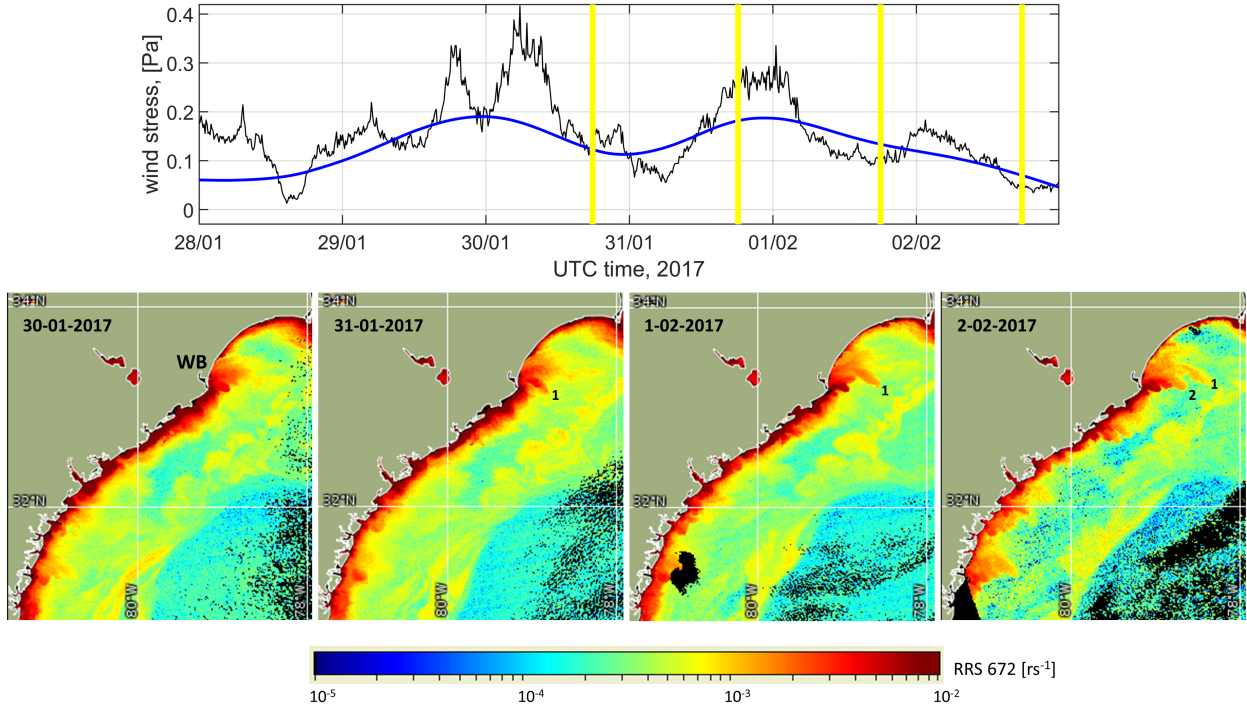


FIG. 15. Cross-shelf WB plume event on April 26, 2019. Top panel: low-pass filtered alongshore wind stress component (heavy line) and instantaneous wind stress magnitude, vertical bar shows time of satellite images; bottom panels: MODIS chlorophyll A from NASA Aqua satellite (left), and VIIRS surface reflectance at 672 nm from NPP satellite (right). Inset shows the location of the study area within the US East Coast. WB is Winyah Bay. Satellite images adapted from NOAA Coastwatch.

shelf plume as described in this paper: an elongated structure detaching from the WB mouth and spreading offshore, crossing almost the entire South Atlantic Bight shelf (Figure 15). Here, the alongshore direction is defined as 40° clockwise from true north. The alongshore low-pass filtered (retaining oscillations longer than 24 h) wind stress component was low, varying within 0.03-0.06 Pa over several days prior to observations, and started to increase just before the images were taken. Since the actual wind is not aligned with the coastline, the instantaneous wind stress magnitude (responsible for mixing) is also shown, and proves that the wind forcing was light to moderate (under 0.1 Pa). We use two proxies for the plume in Figure 15: chlorophyll A measured by a MODIS sensor from the NASA Aqua satellite, and surface reflectance at 672 nm measured by a VIIRS

603 sensor on a NPP satellite as a proxy for the relative amount of sediments (e.g. Stumpf and Pennock
 604 (1989), see also <https://coastwatch.noaa.gov/cwn/index.html>). These measurements
 605 were taken over overlapping time intervals and reveal a similar structure of the WB plume.



606 FIG. 16. Cross-shelf plume event on January 30 - February 2, 2017. Same as in Figure 15, except only
 607 reflectance at 672 nm is shown at the bottom for four days. Numbers 1 and 2 in satellite images refer to
 608 consecutive cross-shelf plumes.

609 A somewhat different plume structure was observed in late January-early February of 2017
 610 (Figure 16). During this time the wind was also upwelling-favorable, but significantly stronger
 611 than in Figure 15: the low-passed alongshore wind stress continuously exceeded 0.1 Pa starting
 612 from the beginning of January 29 through the mid-day of February 2. The instantaneous wind
 613 stress magnitude on several occasions was in the range of 0.3-0.4 Pa. As a result, the WB plume
 614 retained contact with the coastline and the buoyant water was advected both upstream (along
 615 the coast) and offshore, similar to the regime in model run V2W10 (i.e., Figures 7d and 8d).
 616 This diffusive spreading is the most obvious on 1/30/2017, following the strongest wind forcing.

617 The wind subsided in the first half of 1/31, so that a tidal sub-plume extending offshore formed
618 (marked with “1” on the satellite image). Once the buoyant water escaped the nearshore trapping,
619 it continued to develop into a filament-like structure, which can be traced in subsequent images. In
620 addition, a second cross-shelf plume comprising several tidal pulses formed by 2/02 (marked with
621 2) as the low-passed upwelling-favorable wind stress was subsiding over the preceding period of
622 ~ 30 hours. In the event reported in Figure 15, the tidally-averaged freshwater discharge of the Pee
623 Dee River feeding the WB plume was $\sim 850 \text{ m}^3 \text{ s}^{-1}$ prior to observations, while in the event from
624 Figure 16 the Pee Dee River discharge was lower, ranging within $450 - 600 \text{ m}^3 \text{ s}^{-1}$, which could
625 also contribute to more efficient mixing nearshore.

626 Importantly, the formation of a thin detached buoyant layer (which subsequently evolves into a
627 cross-shelf plume) can occur under a variety of inflow conditions at the mouth: when the Burger
628 number is high (e.g., strong nonlinear momentum advection) or low (e.g., geostrophic adjustment
629 is possible), as well as under steady-state or tidally-modulated inflow. This implies two different
630 mechanisms: hydraulically controlled liftoff due to inherent dynamics of the outflow (high Bu) vs
631 thinning of the buoyant layer under the influence of superimposed wind stress (low Bu). The plume
632 maintains its tight transverse structure due to the radiation of internal solitons from the upstream
633 flank in the upwind (downstream) direction. This radiation is primarily due to the advection
634 of momentum associated with the plume and sustained by the ambient wind-driven circulation.
635 Soliton radiation does not require a tidally-pulsating discharge and can operate when the outflow
636 occurs at a constant rate. However, tidally-modulated discharge makes internal wave radiation
637 more robust because each consecutive tidal pulse produces a corresponding soliton.

638 Light wind forcing conditions imply that the wind stress steers the buoyant layer but cannot
639 produce a sufficiently strong shear stress divergence at its base for ambient water entrainment by
640 means of Kelvin-Helmholtz billows. However, observations reported by Yankovsky and Voulgaris
641 (2019) and Yankovsky et al. (2022) indicate that such a plume does undergo mixing, implying
642 alternative mixing mechanisms. Yankovsky and Voulgaris (2019) proposed that the combination
643 of geostrophic and wind-induced shear within the interior fronts can bring the Richardson number
644 below its critical value. This study points to the role of inertial-symmetric instability in driving
645 plume mixing especially within its frontal areas (which include tidal interior fronts). Since the

646 plume remains supercritical, radiating internal solitons can also contribute to mixing, especially as
647 they disintegrate into trains of high frequency internal waves.

648 While the presented experiments identify important mechanisms affecting the cross-shelf plume
649 farther offshore (beyond its near field), even the high-resolution case is not able to resolve their
650 contribution to frontal dynamics. Traditionally, it is perceived that the far field region is less
651 demanding on spatial resolution (Ralston et al. 2017). However, high frequency internal waves
652 propagating around the offshore rim of the plume or radiating into the upwind front are likely to
653 undergo breaking and generate mixing and entrainment. Frequently observed thermal stratification
654 outside of the plume can also cause some leakage of the internal wave energy outside of the plume,
655 although high frequency internal waves are likely to become evanescent outside of the plume due
656 to the lower buoyancy frequency. Resolving internal wave transformation and dissipation in the
657 plume frontal zone would require a spatial resolution of $O(1\text{ m})$ or even less, and can be addressed
658 in the future either by means of an observational campaign or by LES numerical experiments.

659 The model configuration in this study is highly idealized, so we briefly address the model
660 limitations. The most obvious simplification is the constant A_z . However, a more realistic closure
661 scheme can in fact render model results less relevant to observations under the present model
662 configuration. In particular, a closure scheme is likely to predict a stronger eddy viscosity for the
663 unstratified wind-driven flow in the shallow channel so that the bottom boundary layer can easily
664 interact with the plume. Under observed conditions, the surface and bottom boundary layers were
665 separated, in part due to the presence of previously discharged buoyant water, so that the newly
666 discharged wind-driven plume slid on top of the older one (Yankovsky et al. 2022). We maintained
667 this separation in the model by selecting an appropriate value for A_z , which in fact is close to the
668 observed values in the Winyah Bay plume.

669 A more subtle but important model artifact is the channel configuration, where the no normal
670 flow boundary condition is applied at the offshore boundary. The free surface does not relax
671 offshore to the unperturbed condition. As a result, the cross-shelf gradient of the alongshore
672 geostrophic velocity is reduced, thus reducing the offshore momentum advection that sustains the
673 soliton radiation. Lastly, periodic boundary conditions imply that there is no alongshore pressure
674 gradient force which typically opposes the wind stress direction (e.g., Carton (1984)). In this case
675 the alongshore velocity is larger (since its acceleration is now balanced by the bottom stress only),

676 and the plume is likely to be more downwind-swept in our simulations. That is, more realistic
677 open boundary conditions (here, eastern, northern and southern boundaries) would most likely
678 have enhanced the tendency for cross-shelf plume development.

679 **6. Conclusions**

680 We have presented a series of idealized, high-resolution numerical experiments delineating the
681 emergence, dynamics, and parameter space of the newly-identified cross-shelf regime of a river
682 plume. The cross-shelf regime refers to river plumes that propagate offshore, remaining coherent
683 for tens to over one hundred kilometers (as seen in satellite images) away from an estuarine mouth.
684 We first considered a canonical alongshore plume regime in the absence of wind forcing and found
685 that our model captures its well-studied dynamics. The plume undergoes geostrophic adjustment,
686 forming a mid-field buoyant bulge and a far-field coastal current that propagates alongside the
687 shelf. A new result from this case was the presence of inertial-symmetric instability (ISI) within
688 the plume. We were able to resolve beams of ISI leading to deepening of the plume front and
689 lateral mixing.

690 We then incorporated an upwelling-favorable wind forcing into our model and observed the
691 emergence of the cross-shelf regime. The plume detaches from the coast in a thin buoyant layer
692 that then propagates tens of kilometers offshore. Remarkably, the plume remains highly coherent,
693 with its width not increasing substantially with distance from the mouth. The coherence is attributed
694 to radiation of internal solitons in the offshore part of the plume which propagate from upstream
695 to downstream (upwind) flank of the plume. The solitons carry mass and momentum within the
696 plume and suppress diffusive spreading and widening of the plume with distance offshore.

697 We considered the parameter space over which the cross-shelf plume regime arises by introducing
698 tidally-modulated buoyant inflow, varying the river mouth width (causing the Burger number to
699 change), changing wind forcing magnitude, and testing the influence of a different coastal geometry
700 (a promontory). We find that the regime is robust and may come about through various scenarios.
701 These include: low Bu (wide mouth), high Bu (narrow mouth), steady-state inflows, tidally-
702 modulated inflows, different coastal geometries, and light to moderate upwelling-favorable wind
703 stress. We find that the regime breaks down under high wind forcing conditions when the formation
704 of the thin, detached buoyant layer is suppressed by vertical mixing.

705 We additionally presented an ultra-high resolution case (SHR) analogous to our standard case
706 with tidal forcing (S). The dynamics in SHR were consistent with S, albeit permitting the additional
707 resolution of internal wave trains evolving from solitons. The convergence between S and SHR
708 further validated our model setup. In SHR the horizontal resolution was 15 m, allowing us to
709 resolve submesoscale instability dynamics. We observed the presence of large regions of negative
710 Ertel potential vorticity (EPV) at fronts generated between tidal sub-plumes. Vigorous ISI develops
711 out of the negative EPV regions and is seen as alternating beams between the plume fronts. The
712 ISI causes lateral tracer mixing and frontal erosion at the surface, while internal waves and solitons
713 initiate mixing and freshwater penetration vertically.

714 Finally, we looked at observations of the Winyah Bay plume. As predicted by our model, we
715 observe the cross-shelf regime under favorable wind forcing conditions. Overall, the cross-shelf
716 plume appears to be a robust and efficient mechanism for cross-shelf exchange, especially for wide
717 continental shelves affected by significant freshwater discharge. It can be formed on timescales
718 of just a few days, much faster than the classical anticyclonic bulge, and the required forcing
719 conditions are fairly typical, at least for the US East Coast. As a transient feature with a transverse
720 horizontal scale of $O(10\text{ km})$ and depth of just a few meters, cross-shelf plumes are unlikely to be
721 resolved by general circulation or climate models. Hence, their impact on the cross-shelf transport
722 of freshwater with associated suspended and dissolved material of terrigenous origin needs to be
723 parameterized in such models.

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Data availability statement. Observational data are available at the following websites. Satellite images: https://coastwatch.chesapeakebay.noaa.gov/region_cl.php; buoy wind data: <https://www.ndbc.noaa.gov/>; and freshwater discharge: <https://waterdata.usgs.gov/nwis/>. Model data for simulations presented in this study are available upon request. Codes for analysis and figure generation may be accessed at <https://doi.org/10.5281/zenodo.7557654>.

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