Upwelling relaxation and estuarine plumes

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[1] After coastal upwelling, the water properties in the nearshore coastal region close to estuaries is determined by the race between the new estuarine plume traveling along the coast and the upwelled front (a marker for the old upwelled plume and the coastal pycnocline) returning to the coast under downwelling winds. Away from an estuary, downwelling winds can return the upwelled front to the coast bringing less dense water nearshore. Near the estuary, the estuarine plume can arrive along the coast and return less dense water to the nearshore region before the upwelled front returns to the coast. Where the plume brings less dense water to the coast first, the plume keeps the upwelled front from returning to the coast. In this region, only the plume and the anthropogenic input and larvae associated with the plume waters influence the nearshore after upwelling. We quantify the extent of the region where the plume is responsible for bringing less dense water to the nearshore and keeping the upwelled front from returning to the coast after upwelling. We successfully tested our predictions against numerical experiments and field observations of the Chesapeake plume near Duck, North Carolina. We argue that this alongshore region exists for other estuaries where the time-integrated upwelling and downwelling wind stresses are comparable.


1. Introduction

[2] Coastal upwelling is a well-understood mechanism [Lentz, 1992; Austin and Lentz, 2002], but the physics of the upwelled front after the upwelling wind stops has received less attention [Send et al., 1987; Alessi et al., 1996; Dale et al., 2008]. Upwelling‐favorable alongshore winds (which blow with the coast on their left in the Northern Hemisphere) cause offshore Ekman transport in the surface mixed layer (Figure 1). At the coast, this offshore transport can result in a displacement of the coastal pycnocline and, potentially, outcropping to the surface. This outcropping, the upwelled front, separates cold, salty upwelled water near the coast from warm, less‐salty surface water offshore. Near estuaries, the surface coastal waters advected offshore also consist of the estuarine plume. On a shallow, stratified shelf, the upwelled front usually forms after about two inertial periods [Austin and Lentz, 2002]; in this study we will focus on observations where the upwelled front has formed.

[3] After the upwelling winds cease or reverse, the upwelled front moves shoreward as an ageostrophic buoyant gravity current for an inertial period [Csanady, 1971]. When there are no winds, this onshore movement stops after a distance equal to the radius of deformation [Csanady, 1971]. The onshore movement of the upwelled front stops after the Rossby adjustment, i.e., the shoreward ageostrophic pressure gradient is balanced by the Coriolis force resulting in an alongshore, geostrophic flow [Austin and Lentz, 2002]. In the absence of downwelling winds or alongshore pressure gradients, the upwelled front can only return to coast if the offshore position of the front is closer than the radius of deformation. Using wind‐reversal timescales for the east and west coast of US, Austin and Lentz [2002] predicts the upwelled front is usually much farther offshore, so the front does not return to coast as part of the Rossby adjustment process. If and when the upwelling winds reverse to downwelling, the shoreward surface Ekman flow forces the upwelled front toward the coast [Dale et al., 2008]. In the absence of a nearby estuary, the return of the upwelled front to the coast brings back the less dense surface water, initially moved offshore during upwelling, to the coast as the upwelled isopycncals return to their pre‐upwelling state.

[4] However, near estuaries, an estuarine plume can arrive along the coast from the estuary before the return of the upwelled front (Figure 2). The plume propagates along the coast in the direction of a coastal Kelvin wave, the downwave direction, from the source estuary [Garvine, 1999; Fong and Geyer, 2001]. The speed and distribution of the plume can be altered by the alongshore winds [Fong and Geyer, 2001]. As we shall discuss below, the arrival of the plume nearshore keeps the front from returning to the coast. The plume is pushed against the coast during downwelling winds, forcing the plume isopycncals nearly upright and weakening the vertical stratification, similar to Williams et al. [2010]. The weak vertical stratification in the plume...
permits vertical mixing, so the cross-shelf transport in the plume is weak. This is similar to the well-mixed nearshore region, ‘inner shelf’, described by Lentz et al. [1999], where the alongshore wind stress and pressure gradient are balanced by bottom friction. Numerical models have shown that this weak cross-shelf transport in the plume keeps the upwelled front and constituents trapped in it from returning to the coast [Austin and Lentz, 2002]. As a result, where the plume arrives along the coast before the upwelled front, the salinity, larval inhabitants, terrestrial nutrient runoff, and pollutants in that region would be consistent with that of the estuary where the plume originated. Outside this region, where the upwelled front returns to the coast, the water properties would be influenced by both the estuarine plume and the coastal processes that affect the water during upwelling and downwelling.

[5] There have been observations of alongshore plumes arriving first in the nearshore region. Along the east coast of US, Fong et al. [1997], Rennie et al. [1999], and Cudaback and Largier [2001] show freshwater plumes arriving along the Maine coast and the North Carolina coast. Along the west coast, Send et al. [1987] shows a plume of warm water from the San Francisco Bay moving along the northern California coast. However, there have also been observations of an upwelled front first returning to the coast during downwelling. In the east coast, Shanks et al. [2000, 2002], Marmorino et al. [2004], and Shanks and Brink [2005] show the upwelled front returning to the North Carolina coast. In the west coast, Farrell et al. [1991], Miller and Emlet [1997], and Dale et al. [2008] show the upwelled front returning to the California coast and the Oregon coast. What is not clear is where the plume or the upwelled front dominates. In this paper, we show that after upwelling there is a region downwave of an estuary where the plume is responsible for first bringing less dense water nearshore. This region will depend on the race between the alongshore propagating estuarine plume whose properties are primarily set by the estuary, and the shoreward returning upwelled front whose properties are set by the older upwelled low-salinity plumes and coastal processes.

[6] We will focus on the role of the plume as the downwelling winds force the upwelled front shoreward. We derived an estimate of the region where the plume keeps the upwelled front from returning to the coast. This estimate is useful downwave of estuaries such as the Chesapeake Bay and Delaware Bay in the east coast of US or the Columbia River and Puget Sound in the west coast of US. The west coast has long upwelling periods followed by weak winds and a narrower shelf compared to the east coast, and while our focus is on the east coast (North Carolina, Figure 3), we generalize our results to other coastal shelves in the discussion. We discuss where our estimate is applicable and how it will vary with the size and nature of the estuary and coastal regions.

[7] Our work differs from prior studies [e.g., Garvine, 1999; Simpson, 1997] because we estimate the length of the region close to an estuary where the plume can keep the upwelled front from returning to the coast after upwelling; our estimate is not the eventual length of a plume. Field observations show that an upwelled front arriving at the coast first does not stop the alongshore propagation of the plume beyond that point [Rennie et al., 1999; Cudaback and Largier, 2001]. The size of the plume derived by Garvine [1999] and the region of freshwater influence, ROFI, described by Simpson [1997] determines the eventual length of the plume along the coast from the source estuary. Furthermore, our work examines the advection of less dense, surface water into the nearshore region, which differs from the nearshore mixing-restratification studies on the ROFI [Linden and Simpson, 1988; Sharples and Simpson, 1993; Souza and Simpson, 1997; Burchard and Hofmeister, 2008]; in these studies, the investigators examine when tidal, wind, and waves vertically mix the nearshore water column, and the subsequent restratification due to the seaward Rossby adjustment of the mixed water column during periods of low mixing. The dynamics in the ROFI, as discussed in the above literature, can tell us when the plume is vertically mixed during downwelling and our estimate works (i.e., the low cross-shelf transport in the mixed plume can prevent the upwelled front from reaching the coast), and when our estimate will fail.

Figure 1. Cartoon of the upwelling mechanism transporting the upwelled front offshore with a cross-shore sectional view. The black dots indicate the nearshore and offshore moorings.

Figure 2. During wind-reversal, the onshore flow can transport the upwelled front back to the coast. The arrival of less dense water along the coast can be due to the front returning to the coast or the arrival of a plume from the upwave estuary. We predict the alongshore extent of the plume when the front returns to the coast near the southern boundary.
The paper is organized as follows. Section 2 discusses the methods and configurations utilized for analysis of the numerical experiments and the observations. We examine the base case and tracer experiments and we also show that, as discussed in previous literature, the presence of the plume keeps the upwelled front from returning to the coast. In section 3, we use observations at Duck, NC, to identify periods after upwelling when the plume brings less dense water along the coast and when the front, forced by downwelling winds, brings less dense water along the coast. In section 4, we derive a predictor to estimate where the plume arrives along the coast before the returning upwelled front, and we test it against numerical experiments. In section 5, we test our predictions against CoOP field observations. In section 6, we apply our predictor to other estuaries, discuss in which systems our predictions are applicable, and extend our work to other systems.

2. Description of Numerical Model

2.1. Configuration of Numerical Model

The numerical model used in this study is the Regional Ocean Modeling System (ROMS). It is a primitive equation finite difference numerical model [Arakawa and Lamb, 1977]. The vertical momentum balance is hydrostatic and a free surface is included. We define a constant horizontal eddy viscosity $A_H$ of $5.0 \text{ m}^2 \text{ s}^{-1}$ and a horizontal diffusivity $A_H$ of $0 \text{ m}^2 \text{ s}^{-1}$. The background vertical eddy viscosity is $\nu = 1.0 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$. The Coriolis parameter $f = 10^{-4} \text{ s}^{-1}$. The vertical eddy viscosity $K_M$ is computed by the Mellor-Yamada level 2.5 turbulence closure scheme [Mellor and Yamada, 1982] using the non-dimensional stability functions from Galperin et al. [1988] and Kantha and Clayson [1994]. Some studies, e.g., Garvine [1999], Stacey et al. [1999], and Fong and Geyer [2001], have shown that in strongly stratified conditions, the MY2.5 scheme underestimates the vertical eddy viscosity, while in weakly stratified conditions the vertical eddy viscosity is overestimated. Nevertheless, the MY2.5 scheme resolves the mixing accurate to first order, which is what we are interested in. The density is computed by a linear equation of state using a saline contraction coefficient of $7.6 \times 10^{-4}$ and is a function of salinity alone. Temperature is kept constant in the model. A passive tracer is included in the estuarine water. This tracer configuration helps distinguish the estuarine plume from the returning upwelled front.

2.1.1. Model Grids

[10] The model utilizes a sigma coordinate system to resolve the vertical structure. We use 20 sigma levels with closer vertical spacing at the surface and bottom to resolve the boundary layers. The horizontal grid is a finite difference scheme with grid size ranging from 1 km (near northern boundary) to 4 km (near southern boundary). The higher resolution at the northern boundary minimizes the formation and downwave propagation of numerical artifacts into the study area. We also use a higher horizontal resolution in the shallowest part of the domain to resolve the nearshore physics that we are examining. The shallowest region is along the coast (10 m), and the deepest region is along the offshore boundary (110 m). The domain in our base case numerical experiment has 174 points in the cross-shore direction and 179 points in the alongshore direction. The location of the boundaries are southern $y = -315$ km, northern $y = 50$ km, western $x = -110$ km, and eastern $x = 100$ km. The barotropic time step is 9 s, and the baroclinic time step is 180 s. A right-handed ’east coast’ coordinate system is used where $+x$ direction is offshore (‘seaward’), $+y$ direction is northward (‘upwave’), and $+z$ direction is upwards (‘skyward’).

2.1.2. Boundary Conditions

[11] The surface momentum boundary conditions are,

$$K_M \frac{\partial (u, v)}{\partial z} \bigg|_{z=0} = \frac{(\tau^{x,y})}{\rho_0}$$

where $\tau^{x,y}$ are cross-shore and alongshore surface wind stresses. In this study, $\tau^{x,y} = 0 \text{ N m}^{-2}$ while the alongshore wind stress is varied. The bottom momentum boundary condition is a linear bottom drag,

$$K_M \frac{\partial (u, v)}{\partial z} \bigg|_{z=-H} = (\nu u_b, \nu v_b)$$

where the bottom drag coefficient is $r = 3 \times 10^{-4} \text{ m s}^{-1}$ and $(u_b, v_b)$ are the bottom velocity in the cross-shore and alongshore directions respectively. The bottom drag is chosen within the range observed by Lentz et al. [2001] for the North Carolina coast.

[12] The northern open boundary conditions (OBC) are determined from numerical experiments using the same
winds as the 3D model but in a 2D alongshore-uniform topography model. This 2D model is a cross-shore section at the northern boundary of the 3D model and has no alongshore variations. The northern OBC implies that the ocean outside the northern boundary can be approximated as an infinite coast with no alongshore variations in forcings or topography, as described by Gan and Allen [2005] and Pringle and Dever [2009]. The southern and eastern edges are open boundaries with Sommerfeld radiation conditions. This radiation condition has the least reflection at the boundary for the dominant wave mode but the other wave modes have higher reflection. To overcome this, the southern boundary also has a six grid point wide sponge layer where the horizontal viscosity gradually increases in the southward direction. The sponge layer helps dissipate the energy of reflected waves preventing them from propagating along the southern open boundary. The western boundary is the coastal wall with free-slip condition. The free-slip imposes no friction between horizontal boundaries and the flow, and there is no normal flow into the wall. The western boundary also has a wide estuary with a freshwater river input into the model domain. Our model estuary has idealized dimensions similar to wide estuaries such as Chesapeake Bay and Delaware Bay.

2.1.3. Bathymetry

The model domain is an idealized representation of the study area and has a wide estuary connected to a uniform alongshore bathymetry coastal ocean (Figure 4a). The coastal ocean bathymetry (Figure 4b) is given by

\[ H(x) = H_o + \alpha x \quad 0 \leq x \leq L = 100 \times 10^3 \text{m} \]  

(3)

where \( H_o = 10 \text{ m} \) is the coastal wall depth, \( \alpha = 0.001 \) is the bottom slope and \( x \) is the cross-shore distance. The estuary mouth is centered at \([x, y] = [0, 0]\) km with a length of 100 km and a width of 20 km. The estuary has 10 m deep walls with a 11 m deep thalweg (Figure 4c). The dimensions of the estuary are idealized and kept constant to prevent any variations in estuarine mixing. The details of the estuary are not first order important because the physics we are interested in depends on the density and thickness of the plume in the coastal ocean. The variations of these parameters are described later in this section.

2.1.4. Stratification

The coastal ocean is modeled after the summer conditions at Duck, NC (Figure 3) as a two layer system separated by a halocline (thickness of 10 m) centered at a depth of 15 m [Waldorf et al., 1995; Alessi et al., 1996]. These dimensions of the halocline simplify the mixing dynamics at the estuary mouth by preventing the coastal halocline from entering the estuary during the spin up. The initial density field is along-shore uniform and is a function of depth and salinity,

\[ S(z) = 34 - 2 \tanh \left( \frac{z + 15}{3} \right) \text{g kg}^{-1} \]  

(4)
where the salinity is 34 g kg\(^{-1}\) at the halocline center, and \(z\) is the depth in meters (Figure 5a). The salinity of the top layer of the coastal ocean is near 32 g kg\(^{-1}\). We report the numerical model salinity using absolute salinity in units of g kg\(^{-1}\), and the CoOP field observations (old data) are in units of psu, as defined by Millero et al. [2008]. For our purposes, practical salinity and absolute salinity are essentially the same.

### 2.2. External Model Forcings

[15] In our numerical experiments, three components are varied to test our prediction of the region where the plume arrives nearshore first: the duration of the upwelling wind, the downwelling wind stress, and the speed of the plume. The details of the variations of these components are described below. Each numerical experiment starts from an initial condition where the flow is at rest. The experiments run with no winds until steady state, and then a wind-forcing is applied to the model domain. The experiment was run for two weeks after the onset of the wind-forcing. Our analysis will focus on the evolution of the nearshore dynamics during and after the wind-forcing.

#### 2.2.1. Wind Forcing

[16] The surface forcing consists of uniform alongshore wind stress, \(\tau_{sw}\). The wind stress begins 22 days into the numerical model run to allow the plume and exchange fluxes at the estuary mouth to stabilize. The wind-forcing begins as an upwelling alongshore wind stress of \(\tau_{sw}\) lasting for an upwelling period of \(t_{uw}\); this period includes one inertial period (\(t_{f} = \frac{2\pi}{\omega_1}\)) to ramp up and another inertial period to ramp down to zero wind stress (Figure 5b). After the end of the upwelling period, the wind reverses and ramps up to a downwelling wind stress of \(B\tau_{sw}\) over an inertial period. The downwelling wind stress remains at \(B\tau_{sw}\) for the remainder of the model run. \(t_{uw}\) controls the offshore position of the front and is varied from 2.5–5 days. \(B\) is a factor that controls the magnitude of the downwelling wind stress relative to the upwelling wind stress and is varied from 0.25–2. In all numerical experiments, \(\tau_{sw} = 0.1\) Pa, roughly a 8 m s\(^{-1}\) wind speed measured at height of 10 m [Fairall et al., 1996] and the changes in wind-forcing in each numerical experiment is made through \(t_{uw}\) and \(B\).

#### 2.2.2. River Forcing

[17] The freshwater flux enters the model domain at the head of the estuary and is applied evenly over the depth. This inflow initiates at the start of the model simulation and is kept constant with time. The exchange flow at the estuary mouth has stabilized after 22 days. The freshwater inflow is fixed at 2000 m\(^3\) s\(^{-1}\) in all our numerical experiments. The initial salinity along the estuary is,

\[
S(x) = S_{co} + \frac{\partial S}{\partial x} x, \quad -100 \text{ km} \leq x \leq 0 \text{ km}
\]

where the surface salinity of coastal ocean, \(S_{co} = 32\) g kg\(^{-1}\) and the along-estuary salinity gradient, \(\frac{\partial S}{\partial x} = 0.2\) g kg\(^{-1}\) km\(^{-1}\) (\(\frac{\partial S}{\partial x} = 0.16\) kg m\(^{-1}\) km\(^{-1}\)). This base case gradient is the stabilized along-estuary salinity gradient at the end of a 100-day simulation with the river flow alone. This helps shorten the spin up time for the estuary in the numerical model by starting the model close to the steady state. In order to change the plume speed, variations of the along-estuary salinity gradient ranging from 0.05–0.25 g kg\(^{-1}\) km\(^{-1}\) (\(\frac{\partial S}{\partial x} = 0.04–0.20\) kg m\(^{-1}\) km\(^{-1}\)) are used to change the density of the buoyant estuarine plume leaving the estuary. In these density variations, the estuary salinity has reasonably adjusted such that the plume water at the estuary mouth varies between 2–10% over the last week before upwelling.

### 2.3. The Response of the Base Case Numerical Experiment

[18] The base case is configured such that there is 2.5 days of upwelling at a wind stress of \(\tau_{uw} = 0.1\) Pa followed by downwelling wind stress of \(\tau_{dw} = -0.1\) Pa until the end of the experiment. The upwelling duration of 2.5 days is chosen from the range of observed wind-reversal timescale in our study area [Austin and Lentz, 1999], so there is 1 more day of upwelling after the coastal halocline outcrops at the coast. The 32.5 g kg\(^{-1}\) isohaline, at the top of the coastal halocline, is used as a marker for the upwelled front. After

![Figure 5](image-url)
the winds reverse to downwelling, the onshore speed of the front is the depth-averaged onshore velocity of the surface mixed layer. In the numerical model, the mixed layer is determined as the depth where the salinity is 0.1 g kg$^{-1}$ higher than the surface salinity. The return of the upwelled front is associated with downwelling winds, and we define the return of the upwelled front to the coast when the bottom salinity at the coast (near the southern boundary) is fresher than 34 g kg$^{-1}$ and the surface-bottom salinity difference is less than our tolerance of 0.5 g kg$^{-1}$. The 34 g kg$^{-1}$ isohaline is the center of the model halocline returning back to intersect with the coastal wall, similar to the pre-upwelling salinity profile. For the base case configuration, the upwelled front returns to the coast on day 27 of the experiment or 2 days after onset of downwelling winds (marked in Figure 6).

While the front is returning to the coast, the estuarine plume is also propagating alongshore as shown in the plot of alongshore surface salinity at the coastal wall versus experiment model time (Figure 6). The vertical salinity contours indicate the upwelled front, and the alongshore distance-time gradient of the sloping salinity contours indicate the plume speed, $c_T$. We use the sloping 31 g kg$^{-1}$ isohaline to determine the plume speed. The plume speed $c_T$, includes the downwelling wind-induced alongshore flow. The downwelling winds also push the plume against the coastal wall, narrowing the plume width and increasing the plume thickness, so the plume interacts with the coastal bottom slope.

2.4. Base Case Tracer Experiment

A passive tracer is used in our base case numerical experiment to show that the upwelled front remains outside the plume boundary. At the onset of wind-reversal, a tracer is injected in the estuary while the rest of the model domain had no tracer. The cross-section shown in Figure 7, is 60 km downwave of the estuary, in the region where the plume arrives nearshore first. The 32.5 g kg$^{-1}$ isohaline marks the position of the upwelled front. During upwelling, 1 day before wind-reversal, we observe the upwelled front moving offshore (Figure 7a) and replaced by the saltier bottom coastal water. When upwelling winds reverse to downwelling, the upwelled front moves shoreward (Figure 7b), but 1 day after wind-reversal, we can observe the leading edge of the plume arriving nearshore before the upwelled front (Figure 7c).
We observed that 3 days after wind-reversal (Figure 7d), the nearshore region occupied by the estuarine plume still had high tracer, and the returning front remained outside the plume boundary. This is important because the presence of the plume nearshore during downwelling is similar to an inner shelf, which keeps the returning upwelled front from returning to the coast, so the region where the plume arrives nearshore first is not influenced by the returning front.

We determine this region where the plume brings less dense water nearshore first and keeps the upwelled front from returning to the coast during downwelling winds. This alongshore extent of the plume is determined when the upwelled front returns back to the coast near the southern boundary. The surface salinity along the coast shows the alongshore extent of the plume when the upwelled front returns to the coast near the southern boundary (Figure 8a). Closer to the estuary, the plume reaches the nearshore first (Figure 8b) while farther downwelling of the estuary the front returns first (Figure 8c). We determined the base case magnitudes of the extent of the region where the plume arrive first, the offshore position of the upwelled front, onshore speed of the returning front, and the plume speed (listed in Table 1).

2.5. Observational Data

The observations we use to support our work are from the moorings and shipboard conductivity, temperature, and depth (CTD) data from the CoOP program in Duck, North Carolina (NC). The reasons for the selection of this site were the large volume of observations, background literature [Lentz and Largier, 2006, and references therein], and its alongshore bottom topography, which minimizes bathymetry induced alongshore variations (Figure 3). The cross-shore array of moored conductivity and temperature (CT) sensors and current meters measured the hydrography from near-surface to near-bottom depths. The mooring measurements spanned from late August to early December 1994. The central mooring array consisted of moorings at the 4 m, 8 m, 13 m, 21 m, and 26 m isobaths ranging from 0.3 km to 16.5 km offshore. The observational data described in our analysis are from moored surface and bottom CT meters positioned cross-shore at 1.5 km and 5.5 km (known as the D1 and D2 moorings). There are mid depth sensors as well, but the surface and bottom moorings are the most useful in determining when the water column is mixed or stratified. The details are in the works by Alessi et al. [1996] and Lentz and Largier [2006]. The shipboard CTD casts were made along cross-shore transects reaching as far as 50 km offshore. These CTD casts were made during the months of August and October in 1994. These transects are arranged in the downwave direction from the Chesapeake Bay mouth to north of Cape Hatteras (Figure 3), and a full survey took 24 hours. In this time, the alongshore plume advection can travel 50 km, so these observational plots are not a snapshot of the system. The cross-shore sampling period for each transect of the array is much shorter (2–4 hours), so cross-shore advection of any water masses are small compared to

Figure 7. The cross-section of salinity (g kg$^{-1}$) and tracer at 60 km downwave of the estuary. The darker shading shows the high tracer arriving with the plume. The sections are at times (a) 1 day before wind-reversal, (b) onset of wind-reversal, (c) 1 day after wind-reversal, and (d) 3 days after wind-reversal. The 32.5 g kg$^{-1}$ isohaline shows the position of the upwelled front.
the transect length and can be treated as a good snapshot of the cross-sectional hydrography. Details on the shipboard casts are given by Waldorf et al. [1995]. The CoOP observations are old observations, so we report them in psu in line with Millero et al. [2008]. This also makes it easy to compare with prior studies which used CoOP observations. Several investigators [Rennie et al., 1999; Lentz et al., 1999; Shanks et al., 2000; Cudaback and Largier, 2001; Shanks et al., 2002; Lentz et al., 2003; Shanks and Brink, 2005] have identified the Chesapeake plume and the upwelled front in the CoOP observations.

3. Observations

[23] Using the CoOP observations near Chesapeake Bay, we go a step further than prior studies by examining events when the plume arrives nearshore first, and when the returning upwelled front arrives nearshore first. The CoOP observations used were taken from 14 August to 14 October (Figure 9).

[24] If upwelling winds persist long enough, an upwelled front forms where the halocline is displaced to the surface near the coast and then moves offshore forced by the surface Ekman transport. Shoreward of the upwelled front, the difference between the surface and bottom salinities is small [Austin and Lentz, 2002]. As an upwelled front moving offshore passes a mooring location, the water column at that location becomes well mixed.

[25] After upwelling winds stop, two possible situations can occur. In the first situation, the upwelled front can return to the coast before the plume. When this happens, the moorings would show the less dense water in the upwelled front moving shoreward past them. Since there is no source of freshwater offshore, wind-induced mixing causes the salinity of the surface water moved offshore by upwelling winds to increase along with the thickness of the mixed layer [Pollard et al., 1973; Fong and Geyer, 2001]. The increase in the mixed layer thickness can be large enough that when the upwelled front returns past the nearshore mooring, it influences both the surface and bottom salinity. Thus, when the upwelled front returns to the coast, we expect both the nearshore surface and bottom mooring to show the water freshening at the same time.

[26] In the second situation, the plume is the first to arrive nearshore bringing fresher water than the upwelled front.

Table 1. The Base Numerical Experiment Forcings and Results as Described in Section 2.3

<table>
<thead>
<tr>
<th>Results</th>
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<tbody>
<tr>
<td>Upwelling wind stress, $\tau_{UW}$</td>
<td>0.1 N m$^{-2}$</td>
</tr>
<tr>
<td>Upwelling wind duration, $t_{UW}$</td>
<td>2.5 days</td>
</tr>
<tr>
<td>Offshore position of the upwelled front, $W$</td>
<td>10.4 km</td>
</tr>
<tr>
<td>Downwelling wind stress, $\tau_{DW}$</td>
<td>$-0.1$ N m$^{-2}$</td>
</tr>
<tr>
<td>Plume speed (no wind-forcing), $c_T$</td>
<td>0.38 m s$^{-1}$</td>
</tr>
<tr>
<td>Plume speed (under downwelling winds), $c_T$</td>
<td>0.59 m s$^{-1}$</td>
</tr>
<tr>
<td>Onshore speed of front returning to coast, $u_{FW}$</td>
<td>0.05 m s$^{-1}$</td>
</tr>
<tr>
<td>Length scale, $L$</td>
<td>104 km</td>
</tr>
</tbody>
</table>
returning to the coast. The Chesapeake plume usually causes the nearshore surface salinity to freshen by 4–6 psu while the bottom salinity remains mostly undisturbed (except during strong downwelling winds). In addition to the fresher plume water, the blunt wedge shape of the plume \cite{Lentz and Helfrich, 2002} would bring fresh, less dense water at the nearshore mooring before the offshore (Figure 2). With these heuristics, we can analyze the CoOP observations to determine when, after upwelling, the plume arrives nearshore before the upwelled front returns to the coast during downwelling winds.

\[27\] In the CoOP observations, the upwelling winds often reversed before an upwelled front formed. In our analysis of the CoOP observations, we examined events where an upwelled front had formed. We divide these events into groups where upwelling winds reverse or relax. The upwelling wind-relaxation (UWR) events are defined as the upwelling wind stress becoming zero for at least a day. The wind-reversal (WR) events are defined as the upwelling winds reversing to downwelling winds. A suffix is attached to indicate if the plume (P) or the returning front (F) first reaches the nearshore region as determined with the heuristics discussed above, and the number indicates the period; WR2F is the second wind-reversal period and the upwelled front (F) returns nearshore before the plume. In the period when surface and bottom salinity data was being gathered, the CoOP observations contain seven upwelling events where an upwelled front was observed: five of which were followed by wind-reversals and two were followed by wind-relaxation (Figure 9).

\[28\] In three of the five wind-reversal events, saltier water than that moved offshore during upwelling arrives nearshore, and the difference between surface and bottom salinity at the nearshore mooring remained small as the nearshore water column became less salty, suggesting that the upwelled front returns to coast first. The wind-reversal period, WR2F, during 21–24 August is an example of the front returning to the coast before the plume arrives. By middle of 22 August the upwelling winds transported the front offshore forming an inner shelf (Figure 10b). During the downwelling winds on 23 August, we observe saltier water nearshore (Figure 10c) than that moved offshore by upwelling winds (Figure 10b). This is consistent with the upwelled front returning to coast before the fresher plume; the plume arrival was observed a day later, early on 25 August (Figure 10d).

\[29\] In two of the five wind-reversal events, the nearshore surface salinity is much fresher than it would be if the upwelled front returned to the coast, and the nearshore surface salinity was freshening while the bottom salinity was mostly undisturbed giving a salinity difference of about 6 psu. The bottom salinity is undisturbed when the plume

\[27\] Figure 9. The time series plot at the Duck pier station. (a) Low-pass filtered (PL33 filter \cite{Flagg et al., 1976}) alongshore wind stress \(\tau_{sx}\) (Pa). Positive values are upwelling and negative values are downwelling. Near-surface and near-bottom salinity (g kg\(^{-1}\)) time series for moorings at (b) 1.5 km and (c) 5.5 km offshore. The dashed lines are the near-bottom salinity and the solid lines are the near-surface salinity. UWR1 indicates the period of upwelling wind relaxation [UWR] event 1. WR1P indicates the period of wind-reversal [WR] event 1 with plume-[P] arriving before front-[F]. The light shades show the upwelling relaxation duration and the darker shades show the duration of wind-reversal events.
does not narrow and deepen significantly during weak downwelling wind stress. These observations are consistent with the arrival of the plume because the upwelled front returning back to the coast would be saltier due to wind-induced mixing [Pollard et al., 1973]. The wind-reversal period, WR1P, during 14–17 August is an example of the plume arriving before the front returns to the coast. During the middle of 15 August the difference between the surface and bottom salinities is very small at the 1.5 km mooring (Figure 11b) indicating the upwelling winds had transported the front past the mooring. Following the wind-reversal in the evening of 15 August, the nearshore region shows the arrival of much fresher water than that upwelled offshore (Figure 10a). This suggests that the plume arrival brought the fresh, less dense water nearshore before the upwelled front.

[30] In the case where the upwelling winds cease but do not reverse (Figure 11a), UWR1, during 19–21 August, we observed that the nearshore mooring (Figure 11b) and the offshore mooring (Figure 11c) show the arrival of less dense water at the coast. This less dense water arrives at the nearshore mooring one day before the offshore mooring and is fresher than the water initially moved offshore; the minimum nearshore salinity after upwelling was 31 psu (Figure 12a), but after wind-relaxation we observe nearshore salinity of 29 psu (Figure 12b). Since there is no source of freshwater offshore, this indicates that the less dense water at the coast is due to the alongshore propagation of freshwater from an upwave source. This is consistent with what is expected if the plume arrives at these stations and is also consistent with the conclusions of Austin and Lentz [2002] that during relaxation the upwelled front does not return to the coast. For the CoOP observations at Duck, the source of freshwater is the estuarine plume from Chesapeake Bay [Rennie et al., 1999].

[31] The above examples of the wind-reversal and wind-relaxation events that followed the upwelling winds show that either the plume or the returning upwelled front during downwelling winds can bring less dense water nearshore along the coast. This suggests that at any point downwave of an estuary, the arrival of less dense water at the coast depends on the race between the alongshore propagation of the plume and the onshore return of the upwelled front during downwelling winds. In some region close to the estuary, the plume will win the race and is responsible for the less dense water nearshore. The arrival of the plume nearshore can keep the upwelled front from returning to the coast (as discussed in section 2.4). In the following section,

Figure 10. (a–d) The cross-shore plots of salinity (psu) and (e–h) the cross-shore plots of sigma density (kg m$^{-3}$) for the same periods at Duck pier station. The depth is in decibars, and the dark line is $\sigma_t = 22$ kg m$^{-3}$. (a, e) On 16 August, the plume arrives before the front returns to coast. (b, f) The restart of the upwelling winds moved the plume offshore on 21 August. (c, g) After wind-reversal on 23 August, the less dense water nearshore is due to return of the upwelled front. (d, h) The plume arrives on 25 August after the front returned to the nearshore during downwelling winds. The temperature (not shown here) remained uniform during these events. The inverted triangles are cross-shore CTD shipboard cast locations and the dots are the locations of the 1.5 km and 5.5 km moorings of CT meters.
we derive a predictor to estimate the alongshore distance where the plume arrives along the coast first and keeps the upwelled front from returning to the coast.

4. Derivation and Testing of Our Predictor

When upwelling winds reverse, the upwelled front moves toward the coast while the plume propagates downwave from the estuary (Figure 2). First, we derive the time it would take for the front to return to the coast, and then we determine the distance the plume travels alongshore in this time. We define this alongshore distance $L$, as the region where the plume arrives nearshore before the upwelled front returns to the coast under downwelling winds. Then, we test this length scale $L$, our predictor, using numerical experiments.

The offshore position of the upwelled front $W$, is determined by the upwelling Ekman velocity $u_{UW}$, and the duration of surface cross-shore transport $t_{UW}$,

$$ W = u_{UW} t_{UW} = u_{UW} (P_{UW} - t_{outcrop}) $$

where $P_{UW}$ is the duration of the upwelling winds, and $t_{outcrop}$ is the time needed for the divergence in cross-shore wind-driven transport to displace the isopycnals to the surface forming a front [Austin and Lentz, 2002]. The outcropping time is determined using numerical experiments with successively longer durations of upwelling winds. The duration of upwelling $P_{UW}$, are plotted against the offshore positions of the front and extrapolated to find the time when the front is at the coast, $x = 0$ km. This $t_{outcrop}$ is determined to be 1.4 days, consistent with the range determined by Austin and Lentz [2002]. This correction is only valid for our experiment since it may depend on the upwelling wind stress, stratification, the depth of the pycnocline, and the bottom slope [Austin and Lentz, 2002]. However, this correction should be of similar magnitude for most upwelling systems with similar cross-shelf bathymetry, e.g., Northern California shelf [Lentz, 1987]. We assume there are no alongshore variations in the position of the upwelled front due to eddies at the edge of the front [Barth, 1994]. The upwelling surface velocity $u_{UW}$, is given by

$$ u_{UW} = \frac{\tau_{UW}}{\rho_o h_{ml}} $$

where $\tau_{UW}$ is the alongshore upwelling wind stress, $\rho_o = 1025$ kg m$^{-3}$ is our reference density of seawater, and $h_{ml}$ is the mixed layer depth in the coastal ocean. The surface mixed layer is assumed to be a flat slab moving offshore and onshore with a speed given by the Ekman velocity.

After the front has moved offshore, wind-reversal causes the front to move toward the coast under downwelling winds. The time for the front to return to the coast $t_{DW}$, under downwelling wind $\tau_{DW}$, is

$$ t_{DW} = \frac{W}{u_{DW}} + 2\pi \frac{f}{\tau_{DW}} $$
where \( \frac{2\pi}{\tau} \) is the inertial period for the downwelling surface Ekman transport to fully develop and \( u_{UW} \) is the downwelling Ekman velocity (m s\(^{-1}\)) given by

\[
u_{DW} = \frac{\tau_{DFW}}{\rho_0 h_{ml}^\text{f}}.
\]  

\tag{9}

The wind-induced mixing can erode stratification causing the surface mixed layer to deepen and become saltier [Pollard et al., 1973]. The mixed layer depth is determined by the maximum surface mixed layer depth scaling from Pollard et al. [1973] (referred as PRT depth). If the downwelling wind stress is less than or equal to the preceding upwelling wind stress, then the surface mixed layer during the downwelling does not deepen. When the PRT depth is less than the initial mixed layer depth in the numerical model, the wind stress does not deepen the mixed layer. Pollard et al. [1973] notes that with sustained winds the mixed layer will continue to deepen beyond the PRT depth, but the rate of this secondary deepening is expected to be smaller.

In a 2 layer system (our configuration), the PRT depth is found by determining the new mixed layer depth whose velocity is given by Ekman dynamics and whose bulk Richardson number is critical, \( R_i = 1 \). The new mixed layer depth is shallower than the old mixed layer. Pollard et al. [1973] notes that with sustained winds the mixed layer will continue to deepen beyond the PRT depth, but the rate of this secondary deepening is expected to be smaller.

\[
R_i = \frac{g \Delta \rho_{\text{new}} h_{\text{new}}^\text{f}}{\rho_0 u^2},
\]  

\tag{10}

where \( \Delta \rho_{\text{new}} \) is the new mixed layer density, \( \rho_0 \) is the density of the bottom layer (we assume the bottom layer is deep), and \( u \) is the wind velocity. From equation (10), we determined that for all the wind stresses used in our numerical experiments (0.05–0.25 Pa) the new mixed layer depth is shallower than the old mixed layer.

\[
\frac{2\pi}{\tau} = \frac{\tau_{DFW}}{\rho_0 h_{ml}^\text{f}}.
\]  

\tag{11}

\[
\Delta \rho_{\text{new}} = \rho_{\text{new}} - \rho_{\text{bot}}.
\]  

\tag{12}

\[
\rho_{\text{new}} = \rho_{\text{old}} h_{\text{old}}^\text{f} + \rho_{\text{bot}} (h_{\text{new}}^\text{f} - h_{\text{ml}}^\text{old}) h_{\text{ml}}^\text{new},
\]  

\tag{13}

where \( \rho_{\text{old}} \) is the old mixed layer density, \( \rho_{\text{bot}} \) is the bottom layer density (we assume the bottom layer is deep), and \( h_{\text{new}}^\text{f} \) is the new mixed layer depth.

\[
h_{\text{ml}}^\text{new} = \left( \frac{g \Delta \rho_{\text{ml}} h_{\text{ml}}}{\rho_0^2 \rho_0^2} \right)^{\frac{1}{2}}
\]  

\tag{14}

From equation (14), we determined that for all the wind stresses used in our numerical experiments (0.05–0.25 Pa) the new mixed layer depth is shallower than the old mixed layer.
layer. This means we expect little deepening of the mixed layer caused by the shear-induced turbulent mixing as a result of an increase in wind stress.

[36] Next, we determine the alongshore distance propagated by the plume in the time, \( t_{\text{DW}} \) from equation (8), for the upwelled front to return to the coast. In the absence of ambient alongshore flow, the plume travels downwave at the speed \( c_p \), given by

\[
 c_p = \sqrt{g' h_p} \quad (15)
\]

where \( g' = \frac{\Delta \rho}{\rho} g \) is the reduced gravity, \( \Delta \rho \) is the density difference between plume and coastal ocean. \( c_p \) depends on the thickness of the plume at the coastal wall \( h_p \) [Lentz and Helfrich, 2002], however when a detailed cross-shore observation of the plume is available, then using the average plume thickness yields better predictions of the plume speed. We also assume in our theory that the reduced gravity in the plume is conserved. For the range of conditions we investigated, the mixing in the plume is a second order effect, however where mixing processes like tides, winds, and breaking waves are significant, as discussed in the ROFI literature [Linden and Simpson, 1988; Sharples and Simpson, 1993; Souza and Simpson, 1997; Burchard and Hofmeister, 2008], the plume density will erode, reducing the plume speed, thus decreasing the accuracy of our predictions. For simplicity in our theory, the influence of bottom friction on \( c_p \) is not considered; we examine the impact of this in our numerical modeling. Using the range of bottom friction found in Lentz et al. [2001] in our base case experiment, we determined the changes in the plume speed is about 15%, so bottom friction is not a significant factor in our study.

[37] Downwelling winds can increase the speed of the plume by driving an ambient downwave flow, \( \overline{\nu} \). Over the shelf, the alongshore wind stress is balanced by the bottom friction \( r = 3 \times 10^{-4} \text{ m s}^{-1} \), so the alongshore momentum balance during downwelling is,

\[
 \overline{\nu} = \frac{W_{\text{DW}}}{\rho_o r} \quad (16)
\]

This wind-induced depth-averaged alongshore flow \( \overline{\nu} \), [Lentz et al., 1999] assists the plume speed \( c_p \). In addition to the wind-induced alongshore flow, a large scale alongshore pressure gradient can also drive an ambient alongshore flow, \( v_{\text{amb}} \) [Lentz, 2008] (in our numerical experiments, we assume \( v_{\text{amb}} = 0 \text{ m s}^{-1} \)). The plume speed, \( c_T \) is

\[
 c_T = c_p + \overline{\nu} = c_p + \frac{W_{\text{DW}}}{\rho_o r} + v_{\text{amb}} \quad (17)
\]

[38] Using the plume speed, we determine the alongshore distance propagated by the plume in the time it takes for the upwelled front to return to the coast. This length scale \( L \), is the region where the plume first arrives nearshore and keeps the upwelled front from returning to the coast.

\[
 L = t_{\text{DW}} c_T = \left( \frac{W}{U_{\text{DW}}} + \frac{2\pi}{f} \right) (c_p + \overline{\nu}) = \left( \frac{W_{\text{DW}}}{U_{\text{DW}}} + \frac{2\pi}{f} \right) \left( c_p + \frac{W_{\text{DW}}}{\rho_o r} + v_{\text{amb}} \right) \quad (18)
\]

[39] In our derivations above, we neglect the effects of mixing of the plume, but Garvine [1999] found that coastal mixing and the discharge strength of the estuary determine the eventual length of the plume. The Garvine [1999] length scale is the upper limit of all our predictions. In our experiments, all the predictions are smaller than the eventual length of the plume, 425 km, so mixing of the plume in the coast was not a significant factor. In the next sections we test our predictor, equation (18), using numerical experiments and field observations.

4.1. Testing Predictions Against Numerical Model

[40] To test our predictions (equation (18)), the numerical model is used to simulate the range of conditions observable in the typical coastal ocean, described in section 2.2, and determine the downwave distance where the plume arrives nearshore before the upwelled front. The predictions are tested in three sets of model runs (Table 2); in each set, only one parameter is varied while the others are fixed, allowing a rigorous testing of our predictor and its components such as the offshore position of the front, onshore velocity of the front, and the plume speed.

[41] We varied the upwelling wind duration in the range of 2.5–5 days, variations in the downwelling wind stress are in the range of 0.025–0.20 Pa, and the variations in the estuary salinity (hence variations in plume density) relative to ambient coastal salinity are in the range of 5–25 g kg\(^{-1}\). Then the variations in our theory are compared with the results from the numerical experiments (Figures 13 and 14). The predictions of the length scale, \( L \) are compared to the results from the numerical experiments in Figure 13. The prediction of the components of our length scale such as the offshore position of front (inner shelf width), onshore velocity of the front, and the plume speed are compared to the results from the numerical experiments in Figure 14. In the experiment where upwelling winds cease (downwelling wind is zero), the upwelled front remained offshore after upwelling inline with the numerical findings of Austin and Lentz [2002], and the estuarine plume brought less dense water downwave of the estuary, as expected from our predictions.

4.2. Accuracy of the Prediction of Our Length Scale

[42] Next, we tested if our predictor \( L \) is reasonable for a range of variations in upwelling duration, downwelling wind stress, and plume density typical of coastal oceans. We can see in Figure 13 that the relationship between our predictions and the results from the numerical experiment are, as expected, linear and accurate to the first order. The best fit between our predictions of \( L \) and the numerical results is obtained when equation (18) is scaled by a factor of 0.6. To explain the magnitude of this factor, we analyze below the components of our predictor.
4.3. Accuracy of the Predictions of the Offshore Position of the Front

In this experiment, we test the accuracy of our prediction of the offshore position of the front, \( W \), for variations in the upwelling duration, downwelling wind stress, and the plume density. We determined that, as expected, the inner shelf width only changes significantly when the upwelling duration is varied (Figure 14a). The offshore position of the front is not significantly changed when the downwelling wind stress or the plume density is varied. The response of our prediction of the inner shelf width matches well with the numerical model results.

4.4. Accuracy of the Predictions of the Onshore Velocity of the Front

In this experiment, we test the accuracy of our prediction of the onshore velocity of the upwelled front, \( u_{\text{upw}} \), for variations in the upwelling duration, downwelling wind stress, and the plume density. We determined that, as expected, the onshore velocity only changes significantly when the downwelling wind stress is varied (Figure 14b). The onshore velocity is not significantly changed when the upwelling duration or the plume density is varied. The response of our prediction of the onshore velocity of the upwelled front is proportional to the numerical results; however, our predic-

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Figure 13. The predictions of alongshore region \( L, \text{ (km)} \) where the plume arrives nearshore before the upwelled front, compared to numerical model results. The predictions and numerical results are in good agreement. The plot shows variations in the upwelling wind duration from 2.5–5 days, variations in the downwelling wind stress from 0.025–0.2 Pa, and variations in estuary salinity (hence plume density) from 5–25 g kg\(^{-1}\).

Figure 14. Similar to Figure 13. Components of scaling from our theory are compared with numerical model experiments. (a) Offshore position of the upwelled front, (b) onshore velocity of the upwelled front, (c) plume speed using equation (15) and the plume thickness at coast, and (d) the plume speed using the more complete form of the plume speed from Gill [1982] and the average plume thickness.
time $L$ (m s$^{-1}$), $c_p$ is the speed of the plume under no wind forcing, $\tau$ is the wind-induced alongshore flow, $V_{am}$ is the ambient alongshore flow, and $L$ (equation (18)) is the alongshore region where the plume arrives first. The last three columns show the mechanism predicted by $L$ to arrive nearshore at Duck, 85 km downwave of Chesapeake Bay, the mechanism observed in the CoOP observations, and if predicted and observed mechanisms match. PL indicates the plume arrives nearshore first and FR indicates the returning upwelled front is first.

When $L > 85$ km, then the plume is predicted to arrive at Duck first, otherwise the front returns to coast first. $\bar{w}$ is the time-averaged wind stress over the duration of the upwelling winds, $t_{UW}$ is the duration of the upwelling winds, $\bar{w}_{DW}$ is the time-averaged downwelling wind stress, and $h$ is the cross-shore averaged plume thickness. $c_p$ is the speed of the plume under no wind forcing, $\tau$ is the wind-induced alongshore flow, $V_{am}$ is the ambient alongshore flow, and $L$ (equation (18)) is the alongshore region where the plume arrives first. The last three columns show the mechanism predicted by $L$ to arrive nearshore at Duck, 85 km downwave of Chesapeake Bay, the mechanism observed in the CoOP observations, and if predicted and observed mechanisms match. PL indicates the plume arrives nearshore first and FR indicates the returning upwelled front is first.

### 4.5. Accuracy of the Predictions of Alongshore Plume Speed

In this experiment, we test the accuracy of our prediction of the plume speed, $c_p$, for variations in the upwelling duration, downwelling wind stress, and the plume density. We determined that, as expected, the plume speed changes significantly when the plume density or the downwelling wind stress is varied (Figure 14c). The plume speed is not significantly changed when the upwelling wind duration is varied; the small variation of plume speed with upwelling duration is due to the estuarine plume becoming saltier (and slower) over the course of the upwelling and downwelling wind event due to wind-induced mixing in the estuary. The response of the plume speed is influenced by downwelling winds because the downwelling winds determine the wind-induced alongshore flow, $\tau$. The response of our prediction of the plume speed is proportional to the numerical results; however, our predictions are larger than the numerical results (Figure 14c).

When we use a more complete form of the phase speed of an interfacial wave in a two layer fluid [Gill, 1982, p. 122] and the average plume thickness in our theory instead of the plume thickness at the coast, our predictions of the plume speed improves significantly (Figure 14d). The average plume thickness can be determined from the cross-shore geometry of the plume. These changes give a more accurate plume speed, however, the average plume thickness is difficult to determine without detailed cross-shore field observations, and the Gill [1982] scaling is more complex and bulky. For simplicity in our prediction of plume speed, we use equation (15) and plume thickness at the coast. If detailed cross-shore plume observations are available then using the average plume depth yields better predictions. Our assumption of no mixing in the estuarine plume also overestimates the plume speed and can explain some of the second order difference in speed between the numerical model and our predictions, shown in Figures 14c and 14d. Another factor that contributes to this second order difference is the increase in plume speed due to the intensification of the cross-shore density gradient due to downwelling winds ($\approx 10$–$20\%$ of $c_p$).

Our predictor, equation (18), agrees reasonably with the numerical experiment results for the range of variations in each of its components. However, our theory makes several simplifying assumptions discussed in section 4, so some divergences between our prediction and the numerical model are expected. Next, we test our predictions against observational data from the Chesapeake Bay, and our predictor is applied to observational data from other estuaries to determine the region where the plume arrives nearshore first.

### 5. Testing Our Predictor Against Field Observations

We predict when the plume brings less dense water nearshore before the returning front returns to the coast after upwelling winds. Then, we compare our predictions with what actually happens in the CoOP observations at Duck, which is 85 km downwave of the Chesapeake Bay. We selected periods in the observations when the nearshore mooring shows that an upwelled front has formed, moved offshore, and then the upwelling winds reversed or ceased. We find seven such periods from the observations at Duck (Figure 9).

We predict where the plume arrives first along the coast using the time-averaged upwelling wind stress $\bar{w}_{UW}$, the upwelling wind duration $t_{UW}$, the time-averaged downwelling wind stress $\bar{w}_{DW}$, the cross-section averaged plume thickness $h$, the ambient alongshore coastal flow $V_{am}$, the downwave plume speed $c_{pw}$, and the wind-induced alongshore flow $\tau$ during each of the seven periods (Table 3).

The time-averaged upwelling wind stress, $\bar{w}_{UW}$, is the time-averaged wind stress over the duration of the upwelling winds; the time-averaged downwelling wind stress, $\bar{w}_{DW}$, is determined similarly. The duration of the upwelling winds, $t_{UW}$, is from the start of upwelling winds until the winds reverse. The cross-section averaged plume thickness is determined using the plume thickness at the coastal wall and a plume width of 10 km, as determined by Rennie et al. [1999] for the CoOP observations. The speed of the plume is determined using the reduced gravity and average thickness of the plume (equation (15)). The wind-induced alongshore flow, $\tau$, is determined by equation (16) using the time-averaged wind stress over the duration of the upwelling winds, $t_{UW}$, and the time-averaged downwelling wind stress, $\bar{w}_{DW}$.
averaged downwelling wind stress and the bottom friction \( r = 3 \times 10^{-3} \text{ m} \text{s}^{-1} \), in the range observed for North Carolina [Lentz et al., 2001]. The ambient coastal flow near Chesapeake Bay is about 0.04 m s\(^{-1}\) in the southward direction [Valle-Levinson and Lwiza, 1997].

[51] When the plume extent predicted by equation (18) is more than the 85 km from Chesapeake Bay to Duck, we expect the plume to be the first to arrive along the coast at Duck, bringing less dense water nearshore, otherwise the returning front reaches to the nearshore first. Next, we determined when the observations at Duck show the plume or the upwelled front arriving first nearshore for each of the seven periods. When the plume arrives first, we expect the nearshore water to be much fresher than the upwelled front returning to the coast. When the front returns to the coast first, we expect that the nearshore water is much saltier than that moved offshore during upwelling. These heuristics are described in detail in section 3. To see if our prediction is successful, we compare the mechanism bringing less dense water nearshore in our observations with the mechanism predicted by equation (18). For each of the seven wind-reversal periods, shown in Table 3, the agreement between the observations and the predictions are good.

6. Discussion

[52] Where the time-integrated upwelling and downwelling wind stresses are comparable, such as the east coast of US, there exists a region downwave of an estuary where the arrival of the plume in the nearshore region prevents the upwelled front from returning to the coast. This happens because the vertical stratification in the plume weakens during downwelling winds and allows vertical mixing, so the weak cross-shore flow in the plume stops the front from returning to the coast. Close to the estuary, the plume arrives nearshore before the upwelled front, but far from the estuary, the upwelled front returns nearshore before the plume. We predicted the size of this region and successfully tested it against numerical experiments and field observations.

[53] After upwelling has moved the surface coastal waters offshore, close to the estuary the plume always arrives nearshore and replaces the nearshore coastal water. The arrival of the plume blocks the upwelled surface coastal water from returning to the coast. Conversely, outside of the region the upwelled surface coastal water can return to the coast bringing back the pollutants, larvae, and nutrients that were moved offshore by upwelling. Thus, where the plume returns to the coast before the upwelled front, the salinity, temperature, larval inhabitants and the pollution next to the shelf are governed by water discharged from the estuary. Outside of this region, the properties of the nearshore waters are governed both by the estuarine discharge and the processes that alter the water as it moves along the coast, offshore during upwelling and onshore during downwelling.

6.1. Limits to our Theory

[54] In regions where the time-integrated downwelling wind stress is significantly less than the time-integrated upwelling wind stress, the upwelled front will not return to the coast. Some portions of the west coast of US have such regions where long periods of upwelling are separated by wind relaxation and weak downwelling winds [Send et al., 1987]. In these regions, the plume propagates downwave until the next period of upwelling winds detaches the plume from the coast and moves it offshore, where it remains. Mixing processes like tides and breaking waves can (i) significantly deepen the surface mixed layer beyond the shear-induced turbulent mixed layer depth, and (ii) erode the plume density. This can decrease the accuracy of our predictions, and furthermore, where these mixing processes are dominant, the nearshore water properties will be strongly dependent on the mixing, and this is when the ROFI literature can help us understand these nearshore dynamics in more detail [Linden and Simpson, 1988; Sharples and Simpson, 1993; Souza and Simpson, 1997; Burchard and Hofmeister, 2008].

6.2. Relation to Prior Work and Extension to Other Estuaries

[55] Our work provides new insight on results of prior studies like Shanks et al. [2002]. Shanks et al. [2002] observed a cluster of estuarine organisms that entered the coastal region with the Chesapeake plume, but farther downwave they appeared to move from the plume to the coastal water. The region where the organisms appear to cross water masses is in the region where the plume arrives nearshore first and keeps the upwelled front from the coast; as a result, the organisms in the front, some of which are the same species of estuarine organisms as in the plume, passively remain offshore outside the plume boundary. This gives the appearance of the estuarine organisms in the plume actively crossing into the adjacent coastal water mass.

[56] The comparisons to field observations and modeling above have been based on the Chesapeake estuary; however, estuaries will govern similar regions in other coastal oceans where there are comparable cycles of time-integrated upwelling and downwelling winds. The size of this region is estimated using the observed values of upwelling wind stress, upwelling duration, the plume speed, downwelling wind stress and the ambient alongshore flow. The monthly average of these parameters were used to determine the mean alongshore extent of the plume in summer and winter (Table 4). The plume speed observations are from Yankovsky and Chapman [1997]; the thickness of plume are from Hickey et al. [1998]. Pettigrew et al. [1998], and Banas et al. [2009]; the ambient alongshore flow observations for the east coast are by Valle-Levinson and Lwiza [1997] and Lentz [2008] and for the west coast are by Berdeau et al. [2002]; and the rest of the parameters are derived from NDBC stations listed in Table 4. We defined the mean upwelling wind duration, \( \tau_{UW} \), as the average duration for all upwelling winds (greater than an inertial period) minus the time for an upwelled front to form (estimated as 1.4 days in our numerical experiments). The mean upwelling wind stress, \( \tau_{UW} \), is the average alongshore wind stress for all upwelling winds greater than an inertial period (similarly for downwelling, \( \tau_{DW} \)). The plume speed is the sum of \( \sqrt{g \beta (\text{equation } 15)} \), the alongshore flow \( \tau_{\text{induced by the downwelling winds, and the ambient alongshore flow, } v_{\text{amb}}. \) We also make the approximation that the surface mixed layer thickness remains constant during the upwelling and downwelling periods.

[57] The predictions of the average alongshore extent of the plume near the east coast and west coast estuaries is in the range of 45–180 km (Table 4). The alongshore extent of
the plume tends to be longer during winter compared to the summer. This was mainly due to relatively higher mean values of the upwelling wind duration, upwelling wind stress, and plume speed during winter. The above parameters are larger during winter, so the upwelled front moves farther offshore during upwelling, and the plume moves farther downwave in the time it takes for the front to return to the coast; this is seen in large estuaries, e.g., Delaware Bay and in smaller systems, e.g., Penobscot River.

### 6.3. Conclusion

Close to an estuary, in the direction of a Kelvin wave, the nearshore hydrography, chemistry, and biology are controlled by the properties of the water leaving the estuary. We find the spatial extent of this region along the coast where the time-integrated upwelling and downwelling wind stresses are of comparable magnitude; this delimits the region most directly under the influence of the estuary, and where both the estuarine and coastal processes influence the nearshore water properties.

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### References


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