ADJUSTMENT OF A COASTAL BUOYANT OUTFLOW UNDER TIDAL AND WIND FORCING

by

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ABSTRACT

The discharge of brackish water from estuaries typically forms distinct coastal plumes, often visible through their color signature due to sediment load and particulate matter from rivers. The processes of mixing and dispersion of coastal plumes are subjected to natural variations in the magnitude and timing of freshwater inflows, tides and meteorological conditions. This study presents shipboard observation of the bulge region of a buoyant plume off Winyah Bay, SC. The observation comprises downward looking 600 and 1200 kHz ADCP (Acoustic Doppler Current Profiler) and CTD measurements. Along with standard CTD casts, water samples were collected to analyze and determine mass sediment concentration using standard filtering techniques. Auxiliary data such as wind, river discharge and coastal sea level measurements were collected from WeatherFlow, USGS streamflow and NOAA tide gauge stations, respectively. The study addresses the evolution of bulge region of the plume under the conditions of low freshwater discharge and light wind forcing. The study also examines the fate of the suspended sediments under high discharge condition. The spatial and temporal salinity and temperature structures from CTD measurements are analyzed to inspect the evolution of the bulge region. The impact of suspended sediments on modifying the density anomaly within the plume is also assessed. Gradient Richardson number is calculated to examine the influence of suspended sediments on the mixing processes within the plume. Analysis of the observations demonstrate that the buoyant water was not dispersed by the wind forcing but formed a well pronounced baroclinic jet with associated front. The
buoyant outflow occurring at semidiurnal tidal frequency first propagated northward with the wind-driven currents, but then turned anticyclonically and continued southward, against the wind-driven current. However, this baroclinic jet never reached the coastline to form a coastal current. Due to wind forcing, the frontal zone width exceeded the baroclinic Rossby radius, and in some cases multiple frontal structures were observed. Overall, the results demonstrate that under certain forcing conditions all buoyant outflow is deflected into a growing bulge and ultimately spreads offshore contributing to the cross-shelf exchange processes.
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CHAPTER 1
INTRODUCTION

The discharge of rivers and estuaries delivered into the coastal ocean often forms coastal plumes with a sharp density front separating the buoyant inflow water from the denser shelf water. The structure of the plume is determined by various factors such as ambient flow, bottom topography, inflow properties and wind forcing. River plumes, created by localized sources of buoyancy along the coastline, show two distinguishable characteristics. One is the formation of a coastal boundary current turning anticyclonically (to the right in the Northern Hemisphere) from the perspective of an observer at the river mouth looking seaward (Beardsley and Hart 1978; Minato 1983). The other is the development of an anticyclonic eddy (often referred to as the bulge) in front of the river mouth. Chao and Boicourt (1986) defined ‘bulge’ as the transition region which separates the initial estuarine outflow from the downstream coastal current (in the sense of Kelvin wave propagation) due to Earth’s rotation. Their numerical model experiments described the bulge as a non-linear region whose growth is modified by the ratio of gravity current propagation speed within the estuary and that along the coast. Bulge formation requires that the estuary width is narrow relative to the Rossby deformation radius, $Rd_i$ (a length scale at which rotational effects become as important as buoyancy or gravity wave effects in the evolution of the flow) (Garvine 1995, Huq 2009). When present, the bulge is a subtidal feature that grows continuously in time, accumulating a significant portion of the river’s freshwater discharge and causing a corresponding reduction in the transport of
freshwater away from the river mouth region in the coastal current (Fong & Geyer 2002, Horner-Devine et al. 2006). Most numerical model experiments have reproduced the anticyclonic non-linear circulation near the river or estuary mouth (e.g., Oey and Mellor 1993, Kourafalou et al. 1996, Garvine 2001, Fong and Geyer 2002). Garvine (1987) imposed an ambient current in the same direction as the coastal current which augments the alongshore transport in the plume, arrests the offshore expansion of the bulge and causes the plume to be steady (Fong and Geyer 2002).

Chao and Boicourt (1986), Chao (1988), Oey and Mellor (1993), and Kourafalou et al. (1996) had used numerical models to study idealized river discharges through estuaries. They observed that the plume typically extends well offshore as a bulge that is attached to the estuary mouth and has anticyclonic flow within. In most cases, a substantially narrower buoyant coastal current appears adjacent to the coast, originating from the plume and propagating with the coast on its right. The steady bulge circulation is considered analytically in Yankovsky and Chapman (1997). For plumes that are detached from the bottom, they found that the flow regime in the bulge is anticyclonic and is in the gradient wind balance. They derived a length scale that represents the extent to which a plume may spread at the surface if buoyant layer depth does not increase, which is $4Rd_i$. They also derived another length scale that characterizes the equilibrium depth $h_b$, from where the frontal isopycnals outcrop to the surface over $Rd_i$. Equilibrium depth $h_b = \sqrt{\frac{20}{g'}}$ where, $g' = g \frac{\Delta \rho}{\rho_0}$ is the reduced gravity associated with the buoyant layer, $\Delta \rho$ is the density anomaly between buoyant plume layer and ambient water, $\rho_0$ is the density, $g$ is the acceleration due to gravity, $Q$ is the inflow discharge and $f$ is the Coriolis parameter.
The rotational adjustment of the free baroclinic jet entering coastal ocean from an estuarine mouth was addressed by Garvine (1987), Avicola and Huq (2003a, 2003b) and Horner-Devine et al. (2015). Avicola and Huq (2003a) used a scaled estuary discharge on a rotating table and reproduced an unsteady, anticyclonic bulge circulation. They observed that as the initial flow exits the bay, the buoyant outflow moves offshore (across-shelf). However, the Coriolis force begins to turn the buoyant intrusion to the right (Northern hemisphere). Subsequently, the buoyant outflow regains communication with the coast and is observed to impact the coastal wall. Afterwards, a recirculating bulge forms in addition to the downshelf coastal current. They scale the bulge depth using a geostrophic scale based on inflow discharge $Q$ and the bulge width using an internal Rossby radius for the bulge based on the geostrophic depth, $L_b = \sqrt{\frac{2Qg'}{f^3}}$.

Horner-Devine et al. (2006) simulated the river plume in the laboratory with a direct inflow condition. They found that the dynamics of the bulge depend on the bulge radius $r_b$ and the offshore displacement of the bulge center $y_c$. Although both $r_b$ and $y_c$ increase with time, they scale with the internal Rossby radius, $L_i$, and inertial radius, $L_i$, respectively which are constant in time. They observed that the degree to which the bulge is pressed against the coast depends on the ratio of the two length scales, $L^* = \frac{L_i}{L_b}$. When $L^* \ll 1$, the bulge is pressed close to the coast relative to its radius and a large fraction of the river discharge flows away from the river mouth in the coastal current. When $L^* \rightarrow 1$, the bulge is forced offshore by the strength of the river inflow, retaining a large fraction of the discharge and reducing the amount of freshwater carried away by the coastal current.
All the aforementioned studies deal with the unforced coastal plume, although the wind stress is frequently important. Whitney and Garvine (2005) quantified the comparative effects of the wind vs buoyancy forcing to determine whether the coastal current is predominantly wind or buoyancy driven. They developed two scales to characterize the wind effects on the along-shelf and across-shelf structure of the plume. They observed that during light winds, large-scale plumes flow down shelf as slender buoyancy-driven coastal currents. Downwelling favorable winds augment this downstream buoyancy driven flow, narrow the plume and mix the water column whereas, the upwelling favorable winds counter the buoyancy driven flow, spread the plume waters offshore and rapidly mix buoyant waters. In upwelling conditions, persistent winds will eventually distort isopycnals and advect buoyant waters offshore in a mixed layer. However, even when the wind stress is light, it still can produce significant effect on the coastal buoyancy-driven current, for instance by steepening the isopycnals and triggering baroclinic instabilities (e.g., Rogers-Cotrone et al. 2008). More recently, Yankovsky and Voulgaris 2019 (hereinafter referred to YV 19) addressed the effect of light upwelling-favorable wind on the bulge region of the Winyah Bay plume. The role of the upwelling-favorable wind on the offshore transport and mixing of buoyant water was extensively studied previously but for the far field of a plume (e.g., Fong and Geyer 2001; Lentz 2004; Houghton et al. 2004). Fong and Geyer (2001) (hereinafter referred to FG 01) observed upwelling winds make the plume wider, thinner, and eventually detach it from the coast, whereas downwelling winds compress the plume against the coast, causing the plume to narrow and thicken. These results are not directly applicable to the source region of a plume because
the buoyant discharge prevents the lateral homogenization of the buoyant layer (postulated by FG 01 and Lentz 2004).

Buoyant outflow from estuaries is also often controlled by tides and occurs at tidal frequencies. Prior studies, such as Garvine and Monk (1974), found that tidal plume fronts in the near field are dynamically like nonrotating gravity current fronts described using experimental and theoretical gravity current models (Benjamin 1968; Britter & Simpson 1978). Horner-Devine et al. (2015) reported that the tidal outflow from an estuary initially occurs as a gravity current and forms a tidal front near the mouth. These tidal fronts can be advected alongshore and/or cross-shore by shelf currents (e.g., Rijnsburger et al. 2018; YV 19) and subsequently can undergo the geostrophic adjustment (YV 19). Rijnsburger et al. (2018) observed that downwelling winds accelerate the fronts, presumably by advecting the front towards and along the coast, consistent with Ekman wind driven transport.

The fate of suspended sediments from the river mouths can be explained in four ways: supply via plumes; initial deposition; resuspension and transport by marine processes; and long-term net accumulation. The first way occurs in this fashion: Immediately after leaving the confines of a river channel, sediment-laden buoyant plume expands either offshore or alongshore, undergoes advection and mixing by physical-oceanographic processes, and drop a portion of their sediment resulting in the initial deposition. Studies of the Jaba Delta of Bougainville, Papua New Guinea (Wright et al. 1980), and the Porong Delta of Indonesia (Hoekstra 1988) concluded that the most rapid deposition takes place immediately off the mouths of shallow rivers that transport sandy loads and is related to the rapid spreading and deceleration of turbulent jets. Wright et al. (1980) supported the studies by indicating that wave breaking near the river mouth
enhances mixing and momentum exchange between buoyant effluents and the seawater and promotes more rapid initial deposition close to the mouth. In situations where the inner-shelf region remains energetic throughout the year or where high-energy conditions coincide with maximum river discharge, resuspension may take place simultaneously with initial deposition or deposition may be delayed (Adams et al. 1987).

Geyer et al. (2004) reviewed the processes influencing sediment transport as it is transported in turn to estuarine, coastal and marine environments, focusing mainly on those processes influenced by the density differences between the sediment-laden outflow and the ambient waters. They demonstrated the importance of frontal dynamics and wave boundary layer processes in the trapping of fine sediment, leading to highly concentrated suspensions that are dense enough to generate hyperpycnal flows. The association of sediment trapping on the continental shelf with the generation of hyperpycnal flows is an important mechanism for cross-shelf transport of fine sediment, capable of extending the deposition of sediment beyond the range of transport by surface plumes. They reported that although the freshwater signatures of surface plumes extend thousands of km beyond the mouths of large rivers, the extent of significant sediment transport is much more limited, due to settling of sediment out of the plume after it detaches from the bottom boundary layer. The exceptions are coastal mud streams, which may extend for thousands of km from their riverine sources, but only occur in very shallow water where resuspension can maintain significant sediment loads. Nowacki et al. (2012) explored the processes of sediment removal from the Columbia River plume and their spatial and temporal variability in the near-field plume region within 7 km of the river mouth. They observed that sediment transport in the thin, fresh Columbia River plume as observed during high river discharge
and spring tide conditions is characterized by a regime of rapid sediment removal near the river mouth, accompanied by a seaward decay in removal rates farther offshore. They presented a schematic showing potential processes to explain observed sediment removal rates (Figure 10, Nowacki et al. 2012).

This study addresses the question of how the superimposed Ekman dynamics associated with the upwelling-favorable wind impacts the evolution of the bulge area of a plume. The study also addresses the fate of the suspended sediments under high discharge condition. Shipboard observations of a coastal plume off Winyah Bay (WB), SC, a partially mixed estuary in the South-Atlantic Bight are used for the study. The spatial and temporal salinity and temperature are analyzed to address the evolution of the recirculating bulge. The impact of suspended sediment concentration to modify the density anomaly within the plume is also analyzed. Gradient Richardson number is calculated to address the influence of suspended sediment on the mixing processes within the plume.
CHAPTER 2

OBSERVATIONS

In October 2015, South Carolina suffered from a record rainfall resulting in massive erosion and flooding. The shipboard surveys originating from Winyah Bay, SC, a partially mixed estuary in the South Atlantic Bight had been carried out by RV F.G. Walton Smith on October 30-31, 2015 and by RV Savannah on November 10-11, 2015. The surveys provided information about the physical, chemical and optical properties of buoyant freshwater. Both surveys mainly consisted of three transects A, B and C extending from the Winyah Bay mouth offshore. For the first survey, each station on each transect was occupied twice at 6 h time interval to allow phase averaging of semidiurnal tidal species. The first survey was performed in the following order: Aa-Ab-Ba-Ca-Bb-Cb with ‘a’ and ‘b’ referring to the first and second transect occupation, respectively. Subsequently, a repeated sampling over 12 h at stations A2 and A5 was carried out on October 31 to resolve the semidiurnal tidal cycle. The second survey was conducted in the following order: Aa-U-Ab-Ba-S-Bb-C. It proceeded in the outward direction from the mouth along transects A, B and C. During the second survey, only transects A and B were sampled twice for the phase averaging. Several additional stations denoted as U and S were occupied between the transect surveys with the purpose of better distinction of the buoyant plume. The results of the 2nd cruise are presented in YV 19. For both surveys, the observations incorporated downward looking 600 and 1200 kHz Acoustic Doppler Current Profiler (ADCP) and CTD measurements. In the 1st cruise, ADCP was mounted at the side of ship’s hull whereas, for
the 2nd cruise ADCP was mounted at the bottom of the ship’s hull. For the 1st and 2nd cruise, the CTD data were sampled at 0.04 s and 0.25 s respectively and the CTD probe was lowered at 0.2-0.4 m/s but the actual rate of descent slightly fluctuated around these values due to the ship’s roll. Along with standard CTD casts, water samples were collected at each station from 2-3 depths representing the surface boundary layer, the pycnocline, and the bottom boundary layer. Collected water samples were used to analyze and determine the mass sediment concentration using standard filtering techniques. For both surveys, auxiliary data include wind, river discharge and coastal sea level measurements. The sea level data are obtained from the NOAA tide gauge stations: 8662245 Oyster Landing in the North Inlet Estuary (around 17 km northward from Winyah Bay mouth) and 8661070 Springmaid Pier on the exposed coast (around 56 km northeastward from Winyah Bay mouth). The freshwater runoff into WB is estimated by the USGS discharge measurements at station 02135200 Pee Dee river at Hwy 701 near Bucksport, SC and at station 02110704 Waccamaw River at Conway Marina at Conway, SC. The atmospheric forcing measurements were acquired from a tower in the Winyah Bay mouth (YV 19). The wind data at the Winyah Bay mouth were accrued from the commercial weather network WeatherFlow (weatherflow.com).
Figure 2.1: Hydrographic transects of the 1st survey. The circles represent the stations of the survey transects.
CHAPTER 3
DATA ANALYSIS

For the first survey, for each station, ADCP data averaging initiated from the time when ship speed relative to the bottom drops below 0.5 m/s. The ADCP data averaging was carried out for 5 min or until the ship speed relative to the bottom exceeded 1 m/s. Since the ADCP measurements for the first survey did not utilize the bottom track mode, ship’s navigation records from GPS have been inspected. Navigation records provided information on the ground speed magnitude in knots and direction in degrees. These speeds and heading data were used to calculate the eastward and northward components of the ship speed. The eastward and northward components of ADCP velocities had also been generated using ship’s raw ADCP velocity measurements. For the first survey, this averaging process provided high quality data in the uppermost ADCP bin which is at 6.05 m below the surface. The depth of subsequent bins incrementally increased by 2.0 m. The station time was determined as the midpoint in ADCP time averaging and the corresponding coordinates from ship’s GPS were assigned as station coordinates. In addition, to get the depth of each station, navigational depth records of the depth below transducer from the ship’s Echo sounder had been utilized. For the first survey, the transducer depth was 2.0 m and the ADCP blanking distance was 3.05 m.

For the second survey, for each station, ADCP data averaging was carried out in similar manner but on several occasions the threshold of ship’s speed relative to the bottom was slightly raised. This averaging process provided high quality data in the
uppermost ADCP bin which is at 3.05 m below the surface. The depth of subsequent bins incrementally increased by 0.5 m. The transducer depth was 2.35 m and the ADCP blanking distance was 0.45 m (YV 19). For both surveys, the CTD downcast data were averaged in 1 s bins.

Winyah Bay meteorological data provided information on mean wind speed and direction. It was used to calculate the northward (meridional) and eastward (zonal) components of wind speed. This zonal and meridional component of wind speed had been utilized to determine the wind stress ($\tau$) using:

$$\tau = \rho_a C_a W^2$$

where, $C_a$ is the drag coefficient, $\rho_a$ is the air density (~ 1.2 kg m$^{-3}$), and $W$ is the wind speed. $C_a$ is a dimensionless drag coefficient. Following paper by Large and Pond (1981),

$$C_a = 1.2 \times 10^{-3} \text{ for } W \leq 11 \text{ m s}^{-1}$$

$$C_a = (0.49 + 0.065 W) \times 10^{-3} \text{ for } 11 \leq W \leq 25 \text{ m s}^{-1}$$

Freshwater spreading from the Winyah Bay mouth had been characterized by the freshwater layer thickness ($h_f$) which is defined as:

$$h_f = \int_{-D}^{0} \frac{s_r - s}{s_r} \, dz$$

where $s$ is the salinity, $s_r = 34.4$ is the reference salinity (as defined by the offshore CTD measurements during the second cruise), $z$ is the vertical coordinate (positive upward) and $D$ is the total water depth. Salinity was extrapolated to the surface at a constant value of the uppermost CTD bin. For all the stations (except A8, B5 and B8) of transects A, B and C, $h_f$ is phase averaged.
The thermal wind balance (TWB) equation had been used as an approximation for observed vertical shear in the coastal currents and was estimated by applying central differences for the spatial derivatives in the following equation:

\[ \frac{\partial v_g}{\partial z} = - \frac{g}{f \rho} \frac{\partial \rho}{\partial x} \]  

(5)

where, \( V_g \) is the northward geostrophic velocity component and \( \rho \) is the density.

To understand the evolution of frontal flow formed by newly discharged buoyant water in the presence of light wind forcing, the baroclinic velocity \( u_i \) in the tidal plume was estimated. The northward propagation speed can be scaled by using geostrophic across-shelf momentum balance if we assume the buoyant layer outcrops offshore over a horizontal distance of a baroclinic Rossby radius \( R_d \):

\[ u_i = \sqrt{g' h} \]  

(6)

where \( h \) is the buoyant layer depth and \( \Delta \rho \), the density anomaly between buoyant plume layer and ambient water had been related to the salinity anomaly \( \Delta s \) between buoyant layer and ambient water by:

\[ \Delta \rho = \gamma \Delta s \]  

(7)

\( \Delta s \) had been obtained by:

\[ \Delta s = s_r - s_b \]  

(8)

here, \( s_b \) is the salinity of the buoyant layer which had been obtained by:

\[ s = \frac{s_f h_f + s_r (h - h_f)}{h} \]  

(9)

here, \( s_f \) is the salinity of freshwater (taken as 0).
For the South Atlantic Bight (SAB) shelf, the following representative values had been selected: reference salinity, $s_r = 34$, $\gamma = 0.76$, and $\rho_0 = 1020$ kg m$^{-3}$.

To understand the impact of wind forcing applied to the buoyant layer, the depth averaged speed of the Ekman drift ($u_e$) had been calculated as:

$$u_e = \frac{\tau}{\rho_0 f h}$$

where, $\tau$ is the total wind stress ($\tau = \sqrt{\tau_x^2 + \tau_y^2}$). Three estimates for depth averaged speed of the Ekman drift had been obtained by averaging wind stress over different time periods. In the first estimate, wind stress was averaged from the last ebb before the transect survey to the end of the transect survey. In the second and third estimate, the averaging of wind stress started from 6 h and 1 h before the transect survey to the end of transect survey respectively.

The collected seawater samples were filtered in order to measure the weight of the suspended sediments in the water sample. The suspended sediment concentration (SSC) was then calculated by first subtracting the filter weight (~0.132 g) and then diving it by the volume of water filtered. The CTD measurements provided information about the optical properties such as beam attenuation, beam transmission and irradiance of the collected freshwater at different depths for each station. The plume density including the sediment concentration was estimated by using the equation,

$$\rho_p = \rho_{pw} + (\rho_s - \rho_{pw}).C$$

where, $\rho_p$ is the total seawater density including the sediment, $\rho_{pw}$ is the seawater density obtained from the CTD measurements, and $C$ is the volumetric concentration estimated from the mass concentration and particle size. $\rho_s$ represents the density of the sedimentary
particles (for quartz, it was assumed as 2650 kg/m³). The mass concentration (SSC) was used to calculate volumetric concentration, C using the following equation,

\[ C = \left(\frac{1}{\rho_s}\right) \cdot SSC \]  

(12)

Combining the two equations above, the density of the fluid-sediment mixture \( \rho \) (\( S, T, p, \) SSC) was calculated using:

\[ \rho = \rho_{sw} + \left(1 - \frac{\rho_{sw}}{\rho_s}\right) \cdot SSC \]  

(13)

where \( \rho_s \) is the sediment density, and \( \rho_{sw} \) is the sea water density and SSC is the mass concentration.

To understand the impacts of sediments on the mixing of the plume, Gradient Richardson number had been calculated using:

\[ Ri = \left(\frac{g}{\rho}\right) \left(\frac{\rho_z}{u_z^2 + v_z^2}\right) \]  

(14)

where, \( u \) and \( v \) are eastward and northward components of the velocity vector, and subscript refer to the partial derivative with respect to \( z \).
CHAPTER 4
RESULTS

Time series of wind stress components (Fig. 4.1) reveals that the meridional component of wind stress, which is primarily responsible for offshore Ekman transport, was consistently positive from before the survey to the end of the survey of the transects, which is the condition favorable for upwelling. However, during the repeated sampling of stations A2 and A5 the meridional component of wind stress became negative which is the downwelling favorable condition. The magnitude of meridional component of wind stress fluctuated frequently but never exceeded 0.05 Pa before and throughout the whole survey which implies the condition of light wind forcing.

The freshwater discharge (Fig. 4.2) depicts that the discharge was low during the survey. The Pee Dee river discharge remained consistent from before the survey (Oct 28) to during the survey and its tidally averaged value was ~300 m$^3$/s while the Waccamaw river discharge was ~125 m$^3$/s. The freshwater input into WB is higher due to the contribution of other tributaries (Kim and Voulgaris, 2005) although Pee Dee river discharge has the largest fraction.

Temporal evolution of the plume can be observed by two consecutive salinity transects Aa and Ab. The cross-shore salinity of transect Aa (Fig. 4.3) exhibits that the first three stations of transect Aa are continuously stratified with salinity. The near-surface salinity is highest in the middle of the transect Aa and decreases both inshore and
The cross-shore temperature of transect Aa (Fig. 4.3) displays lower temperature close to the mouth and a gradual increase of temperature at the seaward stations of transect Aa. The cross-shore salinity of transect Ab exhibits that the first station of transect Ab (Fig. 4.4) has a 2 m-deep low-salinity buoyant layer whereas the seaward stations like 3rd and 6th station have deeper (~ 2.5 m and 3.5 m respectively) low salinity buoyant layer.

The cross-shore temperature profile of transect Ab (Fig. 4.4) reveals that the lowest temperatures are associated with the low salinity buoyant layers. Frontal structure with strong horizontal salinity gradients can be observed at the seaward station (6th station) of transect Ab. Low salinity waters at nearshore stations (1st and 3rd station) represents individual tidal fronts which are the part of the broad frontal zone. The cross-shore salinity of transect Ba (Fig. 4.5) suggests that no low salinity water is found at the 1st station of the transect while the lowest salinity signatures are observed at the seaward stations (from 3rd to 8th station). Low salinity at the last seaward station proves that the plume extended radially from the mouth over more than 12 km. The cross-shore salinity of transect Bb (Fig. 4.5) conveys that no low salinity water is found at the 1st station of the transect while the lowest salinity signature is observed at the seaward station (4th station) like transect Ba. The cross-shore salinity of transect Ca (Fig. 4.5) conveys similar scenario like transect Ba and Bb where no low salinity water is found at first three stations of transect Ca while the lowest salinity is observed at the seaward station (6th station).

The lowest salinity buoyant layer observed at the 1st station of transect Aa and at the 1st, 3rd and 6th station Ab and the presence of high salinity layer at the 2nd station demonstrate that the freshwater layer discharged from the mouth was not advected offshore. Near surface salinity map (Fig. 4.6) reveals that the lowest salinity water was
found at the 1st station of transect Aa and Ab. No low salinity waters are found on the first two stations on transect Ba, Bb, Ca and Cb. Irrespective of that, some near surface low salinity signature is observed at seaward stations of transect Ba (4th, 6th and 8th station) and of transect Ca (5th station). While near surface salinity increases with the increasing distance from the mouth on transect Aa, some seaward stations (3rd to 6th) of transect Ab show low near surface salinity signature. Transect Aa was conducted during the ebb, while Ab was during the flood. So, the buoyant water near the mouth only, seen on transect Aa, represents no offshore advection of the buoyant layer. Instead, the buoyant layer propagates northward away from the mouth.

Spatial distribution of the freshwater layer thickness (Fig. 4.7) discloses that the largest freshwater layer thickness (~1.2 m) is observed at the first station of transect A. However, the first two stations of both transect B and C do not exhibit any freshwater layer thickness which suggests no formation of coastal boundary current. Although some freshwater layer (~ 0.4 m to 0.8 m thickness) are observed at the seaward stations of transect B and C, it may be due to the freshwater input from other sources, possibly from Santee River. Except the 1st station of transect A, high freshwater layer thickness (~1.15 m) can be observed at the seaward station (6th station) of transect A. So, the buoyant water is not dispersed by the wind since the amount of freshwater does not continuously decrease away from the source. The high amount of freshwater in the seaward station suggests that the buoyant outflow executes anticyclonic turning as it undergoes geostrophic adjustment.

Both depth and phase averaged ADCP velocities (Fig. 4.8) show strong northward propagating barotropic current. Nevertheless, the buoyant outflow executing anticyclonic turning propagates against the wind-driven current as it reaches the seaward stations of
transect Ab. So, the wind forcing does not dominate the buoyancy forcing. Although transect Aa shows negative (southward) shear, the strong positive (northward) shear between station 2 and 3 of transect Ab suggests the northward propagation of the buoyant layer (Fig. 4.9).

Time series salinity profiles of station A2 (Fig. 10) unveils the presence of 4 m deep low salinity water (~29.19 psu) about at around 11.45 on Oct 31 while no low salinity water was observed prior to that time. Time series salinity profiles of station A5 (Fig. 10) unveils the presence of 6 m deep low salinity water (~ 28.35 psu) at around 10.30 on Oct 31 while no low salinity water was observed prior to that time. The presence of deeper low salinity water at earlier time on station A5 portray that the northward propagating geostrophic jet circulates back toward station A5 creating an anticyclonic bulge after undergoing rotational adjustment. Afterwards the geostrophic jet approaches towards station A2 as suggested by low salinity signature at later timeframe on station A2.

Time series freshwater layer thickness profile for station A5 (Fig. 4.11) reveals that the largest amount of freshwater layer thickness (~ 2.1 m) is observed at around 10.30 on Oct 31. The larger freshwater layer thickness signature at earlier time on station A5 divulge that the recirculated freshwater discharged from the mouth returns first on station A5. The meridional component of depth averaged ADCP velocity for station A5 never becomes negative throughout the tidal period. Time series freshwater layer thickness profile for station A2 (Fig. 4.11) shows that the largest amount of freshwater layer thickness (~ 1.4 m) is observed at around 16.30 on Oct 31. The larger freshwater layer thickness signature at later time on station A2 reveals that the recirculating water reaches station A5 first and then spreads toward station A2. The time series of meridional component of depth averaged
ADCP velocity for station A2 like station A5 reveals that the meridional component of depth averaged ADCP velocity never went negative throughout the whole tidal period which implies that after the buoyant outflow occurring at semidiurnal tidal frequency executing anticyclonic turning propagates against the wind driven current. The freshwater layer thickness and the meridional component of depth averaged ADCP velocity show positive correlation. So, the highest meridional velocity component corresponds to the largest amount of freshwater layer thickness.

YV 19 proposed a scenario for anticyclonic bulge formation. They mentioned that in the presence of light wind forcing, frontal flow formed by newly discharged buoyant water undergoes a rotational adjustment and ultimately forms a geostrophic jet along the rim of the tidal plume. They also suggested that the formation of this front should not be inhibited by the superimposed wind stress which means the Coriolis force associated with the newly discharged buoyant flow should exceed the wind stress forcing applied to the buoyant layer. Time series of geostrophic and Ekman velocity estimate for station A5 (Fig 4.12) endorses the statement of YV 19 as the baroclinic velocity in the tidal plume exceeding the depth averaged speed of Ekman drift throughout the tidal period. Time series of geostrophic and Ekman velocity estimate for station A2 further supports the statement of YV 19 as throughout the tidal period except 7.00 shows the baroclinic velocity in the tidal plume exceeding the depth averaged speed of Ekman drift. The exception of 7.00 on Oct 31 is due to the low salinity anomaly between the buoyant plume layer and ambient water and small buoyant plume layer depth. The geostrophic and Ekman velocity estimates for transect Aa and Ab validates the statement of YV 19 as well since the baroclinic
velocity in the tidal plume exceeding the depth averaged speed of Ekman drift for all the stations of both transects.

YV 19 also formulated conditions favorable for the formation of interior fronts. They described that interior fronts can occur if Ekman transport separates them in space which requires the wind driven offshore excursion of the buoyant layer on the tidal time scale exceeds the cross-shore scale of the front which is typically associated with internal Rossby radius. They summarized their results in Fig 11 in their paper by stating that for $\Delta s$ values greater than those outlined by the right limit (Fig 11, dashed lines), the individual tidal fronts will be dispersed by the wind stress so that the plume will be homogenized; for $\Delta s$ values smaller than those outlined by the left limit (Fig 11, solid lines), individual tidal fronts will not be separated, rather will be blended into one broad frontal zone. During 2nd survey, the conditions (freshwater discharge ~ 800 m$^3$/s) were quite optimal for the interior front formation when the alongshore wind stress was 0.04-0.05 Pa (YV 19). However, since during the 1st survey, the discharge was low (~ 400 m$^3$/s), the condition was not favorable for well frontal separation. This statement is endorsed by the observation of the cross-shore salinity profiles of transect Ab where multiple well separated tidal fronts were observed whereas at transect Ba the merging of individual tidal fronts into one frontal zone was observed.

For the 2$^{nd}$ survey, the role of suspended sediments is analyzed to check their impact on density distribution and mixing within the plume. The wind during the 2$^{nd}$ survey changed from strong downwelling-favorable (averaging 13.7 m/s) before the survey (9 November) to light upwelling-favorable (~5 m/s) on November 11. Tides in the survey area are predominantly semidiurnal. The tidally averaged freshwater discharge at USGS
station 02135200 was progressively growing prior to the survey and exceeded 750 m/s on November 9 (YV 19).

For all transects, suspended sediment concentration shows a positive correlation with the optical properties of water like beam attenuation, beam transmission and irradiance. The relationship between suspended sediment concentration and beam attenuation also provides information about the highest suspended sediment concentration which is found on transect A. The suspended sediment concentration along transects Aa, Ab and Bb reveal that highest sediment concentration is found near the mouth and at near the bottom. Transect C shows the sediment concentration becoming higher from the surface to the bottom of the 1st station (Fig. 4.13). The cross-shore sediment concentration reveals that suspended sediments from the mouth gets deposited to the bottom at the 1st station without being carried seaward by the buoyant plume water. The cross-shore density anomaly structures of all transects (Fig. 4.14) provide support of the earlier statement as the highest density anomaly is found near the mouth and at the bottom of the station with the exception of transect C where the density anomaly gradually increased with depth at the 1st station. The suspended sediment concentration has a positive correlation with the density anomaly, ergo, the highest suspended sediments correspond to greater density anomaly. The Gradient Richardson number, calculated with actual density by considering the suspended sediment concentration, show no mixing process due to deposition of suspended sediments since higher suspended sediment concentration are well outside of the buoyant plume layer (Fig 4.15).
Figure 4.1: Time series of eastward (x) and northward (y) components of the wind stress (top); astronomical tidal sea level oscillations at Oyster Landing and Springmaid Pier NOAA tide gauge stations (bottom); Periods of individual transects are shown as black bars in the top panel.
Figure 4.2: Instantaneous freshwater discharge measured at USGS 02135200 Pee Dee River at Hwy 701 near Bucksport, SC.
Figure 4.3: Vertical salinity (top) and temperature (bottom) transect Aa. Both measured (dots) and interpolated (contour lines) values are shown.
Figure 4.4: Vertical salinity (top) and temperature (bottom) transect Ab. Both measured (dots) and interpolated (contour lines) values are shown.
Figure 4.5: Vertical salinity transect Ba (top), Bb (middle) and Ca (bottom). Both measured (dots) and interpolated (contour lines) values are shown.

Figure 4.6: Near surface salinity map
Figure 4.7: Spatial distribution of the freshwater layer thickness $h_f$. 
Figure 4.8: Depth-averaged and Phase-averaged ADCP velocities.
Figure 4.9: Cross-shore density and geostrophic shear of transect Aa (top) & Ab (bottom).
Figure 4.10: Time series vertical salinity profiles at station A2 (top) and A5 (bottom).
Figure 4.11: Time series freshwater layer thickness and depth averaged ADCP velocity at station A2 (top) and A5 (bottom).
Figure 4.12: Time series Ekman and geostrophic velocity of station A2 and A5 (top), cross-shore Ekman and geostrophic velocity of transect Aa and Ab (bottom).
Figure 4.13: Cross-shore suspended sediment concentration of transect Aa (top), Ab (middle) and C (bottom). Contour lines represent interpolated density distribution of individual transects.
Figure 4.14: Cross-shore density anomaly of transect Aa (top), Ab (middle) and C (bottom). Contour lines represent interpolated density distribution of individual transects.
Figure 4.15: Gradient Richardson number calculated using actual density (top) and CTD density (bottom).
CHAPTER 5
DISCUSSION

The theoretical predictions and observations of a far field response to upwelling favorable wind are not directly applicable to the bulge region of the plume. Observational studies reveal that alongshore winds that oppose the buoyant coastal current (upwelling-favorable winds) can spread the buoyant freshwater offshore by wind induced Ekman drift as well as can prompt the buoyant water to detach from the coast and eventually disperse if the winds are strong enough. As the plume spreads offshore, its surface area exposed to the wind increases and it becomes more receptive to wind induced mixing. YV 19 observed the formation of a buoyant plume bulge under conditions of high freshwater discharge and light upwelling favorable wind forcing. The wind forcing didn’t suppress the inherent dynamics of the plume which allowed anticyclonic turning and southward flow of the buoyant water. Their observed structure of the bulge differed from theoretical predictions of FG01 and L04, since a coastal source of buoyant discharge interdicted the lateral homogenization of buoyant layer. They observed the increase of plume depth in a step-like fashion rather than continuously. Estimation of gradient Richardson number showed the likelihood of mixing or entrainment not only at the offshore edge of the plume but also in the proximity of the observed steps. They confirmed that these steps represented tidal fronts which underwent geostrophic adjustment and were advected offshore by the superimposed Ekman drift.
This study inquires the evolution of a buoyant plume under low freshwater discharge and upwelling favorable wind condition. Like YV 19, since the wind forcing was light, it didn’t completely inhibit the intrinsic dynamics of the plume. The buoyant water is not dispersed by the wind as the amount of freshwater does not continuously decrease away from the source. Rather than spreading radially offshore, the discharged freshwater from the mouth travelled northwards as a fast-geostrophic jet. The buoyant layer as it advected offshore by the superimposed Ekman drift, underwent geostrophic adjustment. This buoyant outflow occurring at semidiurnal frequency executes anticyclonic turning and propagates against the wind driven current. The buoyant layer first came back at a seaward station (A5) evident from the high freshwater layer thickness at earlier time on station A5. This buoyant layer then approaches towards nearshore station (A2) evident from the high freshwater layer thickness at later time on station A2. However, this baroclinic jet never regains its communication with the coastline evident from the high salinity waters on transect C. It implies that all the buoyant water goes into the bulge and ultimately it spreads offshore rather than alongshore. The structure of the anticyclonic bulge is altered by these dynamics.

YV 19 formulated conditions favorable for interior front formation. They summarized the results for three values of wind stress (0.03, 0.04 and 0.05 Pa) and for a broad range of both salinity anomalies ($\Delta s$) and freshwater discharge ($q_r$). They also noted that strong mixing (low $\Delta s$) is not favorable for interior front formation. Since the $\Delta s$ and $q_r$ values encountered during 1st survey was low, tidal fronts were not well separated by wind advection. At transect Ab, multiple well separated tidal fronts can be observed corresponding to stronger wind events prior that transect survey. However, conditions
during the 2nd survey were quite optimal for the interior front formation when the alongshore wind stress was 0.04-0.05 Pa. At transect C, well separated multiple tidal fronts can be observed (Figure 8, YV 19)

Yankovsky and Chapman (1997) obtained an estimate for the maximum seaward expansion of the surface advected plume based only on the inflow properties. They found that a weak buoyant inflow or larger density difference yields the spreading of a surface-advected a minimum of more than four inflow Rossby radius offshore. is referred as the baroclinic Rossby radius of the buoyant inflow. Density anomaly and buoyant plume layer depth for the 5th station of transect A of the 1st cruise yield the baroclinic Rossby radius as 5.6 km. So, the maximum seaward expansion of the plume becomes about 23.5 km which can be observed on the cross-shore salinity profile of transect Ab where plume extends further offshore than last seaward station.

The fate of suspended sediments from the river mouths can be unfolded in four ways: supply via plumes; initial deposition; resuspension and transport by marine processes; and long-term net accumulation. For the 2nd cruise, the suspended sediment concentration along transects Aa, Ab and Bb revealed that highest sediment concentration is found near the mouth and at near the bottom. The cross-shore sediment concentration showed that suspended sediments from the mouth gets deposited to the bottom at the 1st station without being carried seaward by the buoyant plume water (initial deposition). The suspended sediment concentration showed a positive correlation with the density anomaly and the Gradient Richardson number showed no mixing process due to deposition of suspended sediments since higher suspended sediment concentration are well outside of the buoyant plume layer.
True color satellite image of the northern South Atlantic Bight (SAB) obtained on Oct 30, 2015 (Fig. 5.2) shows the presence of high turbidity waters nearshore (blue color) which is likely associated with the energetic wind event prior to the survey. In addition, the dark (brown-purple) color specifies freshwater with a high concentration of organic matter delivered by coastal plain blackwater rivers (Meyer, 1990). The true color satellite image (Fig. 5.2) exhibits the northward excursion of the buoyant water from the mouth and the execution of anticyclonic turning of the buoyant layer. It shows the buoyant layer coming back at a station far away from the mouth. The Chl-a concentration obtained from MODIS, AQUA on Oct 30, 2015 (Fig. 5.3) shows high Chl-a concentration (~ 8 mg/m³) offshore. The Chl-a concentration obtained from MODIS, AQUA on Oct 31, 2015 provides a clear exhibition of northward excursion of buoyant water ejected from the mouth, anticyclonic turning of the buoyant layer and ultimately spreading offshore. The remote-sensing reflectance (Rrs) is related to the inherent optical properties of the water and its dissolved and particulate constituents. The remote-sensing reflectance in the red wavelengths (672 nm on VIIRS) can be used as a surrogate for sediment concentration in the water column. As a sediment index, Rrs for red wavelengths indicates relative amounts of sediment only (high vs. low sediment amounts) and is not an estimation of sediment concentration or suspended matter concentration. The Rrs 672 obtained from VIIRS, NPP on both Oct 30 and 31, 2015 (Fig. 5.4) shows high sediment amounts (~ 0.005 sr⁻¹) away from the mouth. They reveal the same picture as Chl-a concentration that the buoyant water from the mouth undertake northward excursion, then turns anticyclonically and comes back at offshore stations and ultimately spreads offshore. The diffuse attenuation coefficient in water indicates how strongly light intensity at a specified wavelength is attenuated within
the water column. This parameter has wide applicability in ocean optics, as it is directly related to the presence of scattering particles in the water column, either organic or inorganic, and thus is an indication of water clarity. The Kd 490 from MODIS, AQUA reveals similar picture of formation of semi-circular unenclosed jet. Both Chl-a concentration from MODIS, AQUA and Rrs 672 from VIIRS, NPP exhibit formation of semi-circular jets formed after consecutive tidal periods. Several curved filaments can also be observed from VIIRS, NPP. The observed interior tidal fronts are responsible for the disruption of continuous anticyclonic flow within the bulge and prevention of the formation of downstream buoyancy driven current. They are effective pathways for buoyant water to escape nearshore confinement of the conventional coastal buoyancy driven current. This scenario is consistent with satellite images obtained during both cruises. By the end of 2nd survey, a well pronounced anticyclonic bulge was formed in Winyah Bay with its most exterior front not returning back to the coast just like 1st survey. Two days after the end of the 2nd survey, anticyclonic bulge disintegrated into several vortices detached from the coast.
Figure 5.1: Favorable conditions for interior front formation based on the scaling by YV 19 and shown for three different wind stresses. Shaded area delineates the conditions encountered during the reported surveys off WB.
Figure 5.2: True color satellite image of the northern SAB obtained on Oct 30, 2015.
Figure 5.3: Chl-a concentration from MODIS, AQUA on Oct 30, 2015 (top) and Oct 31, 2015 (bottom)
Figure 5.4: Rrs 672 from VIIRS, NPP on Oct 30, 2015 (top) and Oct 31, 2015 (bottom)
Figure 5.5: Kd-490 from MODIS, AQUA on Oct 31, 2015.
Figure 5.6: True color satellite image from VIIRS, NPP (top), Chl-a concentration from MODIS, AQUA (middle) and Rrs-672 from VIIRS, NPP (bottom) obtained on Nov 11, 2015.
CHAPTER 6
CONCLUSION

This study reviews the observations of a buoyant coastal plume bulge off Winyah Bay, South Carolina which was formed under conditions of low freshwater discharge and upwelling favorable wind forcing. The evolution of an anticyclonic bulge formed by tidally modulated estuarine outflow to the light upwelling favorable wind appears to be more complex than the far field response which had been described by a slab-like model by FG 01. Since the wind condition was light, it did not completely inhibit the inherent dynamics of the plume. Following an ebb cycle, the newly discharged buoyant freshwater from the mouth propagates northward by forming a geostrophic jet. This jet then undergoes a rotational adjustment allowing anticyclonic turning and southward flow of the buoyant water against the wind driven current. This buoyant water first appears at a seaward station, around 7.8 km away from the mouth. Afterwards, it approaches towards nearshore station, around 1.5 km away from the mouth. The buoyant recirculated freshwater gets accrued within the bulge and never regains its communication with the coastline. Ultimately this anticyclonic bulge spreads offshore rather than alongshore. Scaling theory defining parameter range for interior front formation by YV 19 reveals that because of low riverine discharge and tidal plume salinity anomaly under light wind forcing condition, there is little to no frontal separation within the bulge. Mixing and entrainment within the bulge eventuates as a result of the superposition of geostrophic and wind induced shear.
Study of buoyant coastal plumes will aid us to understand the processes how freshwater from estuaries mix with the coastal ocean since it is important for coastal ocean circulation and ecology. Plumes work like an interface where exchange of biogeochemical materials like sediment, nutrients and organic matter takes place and the transport and degradation of these materials fundamentally govern the net flux of chemical constituents such as carbon, nitrogen and trace metals. Study of a coastal plume is necessary to have better understanding of sensitivity of our coastal environment from the detrimental effects like pollution, eutrophication and salt intrusion. In addition, under current conditions of warming climate and accelerating hydrological cycle when flooding events are frequent and intense, a better understanding of freshwater discharge dispersal and mixing in coastal ocean will help to protect our fragile ecosystem.
REFERENCES


