Tidal Mixing Sustains a Bottom-Trapped River Plume and Buoyant Coastal Current on an Energetic Continental Shelf

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Abstract Conventional wisdoms on river plume dynamics suggest that a down-shelf buoyant coastal current will ultimately be trapped at a specific depth, that is, the trapping depth, as constrained by riverine outflow and offshore bottom Ekman transport. Theoretically, a prerequisite down-shelf current is necessary to form a stable bottom-trapped river plume. In this study an alternative is described by carrying out a modeling study on the Zhe-Min Coastal Current (ZMCC). Buoyant water from the Changjiang River is a major factor driving the ZMCC, as is common in bottom-trapped river plumes; however, the trapping depth is more determined by tidal mixing. When the plume water comes to the sloping topography, strong tidal mixing induces a mixing front, shoreward of which the bottom Ekman layer occupies the entire water column. Such a tidal-induced front maintains a down-shelf frontal current, which is intensified both at the surface due to the thermal wind balance and on the top of bottom boundary layer due to the tidal rectification. Direct wind-induced transport only covers a small fraction of the ZMCC; however, it redistributes the plume water and, thus, affects the coastal current. The tide-induced frontal trapping depth varies much less between seasons than that predicted by previous plume theories. Instead, it fluctuates strongly in the spring-neap cycle. Even in summer when upwelling-favorable winds prevail, the mixing front still sustains a down-shelf coastal current. Intense tidal mixing exists in many coastal waters, which might be an alternative mechanism in forming bottom-trapped river plumes and their associated buoyant coastal current.

Plain Language Summary Large rivers, such as the Amazon, Mississippi, and Changjiang, export a huge amount of terrestrial materials to the receiving waterbodies. Typically, the lighter riverine freshwater mixes with the denser ambient seawater and then propagates rightward along the coast (in the Northern Hemisphere), which is known as the buoyant coastal current. Large river-induced buoyant coastal current can propagate hundreds of kilometers. It frequently causes harmful algal blooms, hypoxia, and other environmental problems. Therefore, understanding the formation mechanisms of buoyant coastal currents is very important. Conventional theories on buoyant coastal currents are very successful; however, they rarely consider the effect of tidal mixing, which exists ubiquitously in coastal waters. Here in this study, we investigated this issue by using the Changjiang River-induced buoyant coastal current as an example, through a series of well-designed numerical experiments. The results showed that tidal mixing plays an essential role in maintaining a stable buoyant coastal current under unfavorable wind conditions. Without the tidal mixing, the stable buoyant coastal current can hardly exist. The finding of this study can promote our understandings on the coastal dynamic processes and other relevant processes.

1. Introduction

When river water enters the sea, a buoyant river plume will form, with a sharp density front separating the plume water from the denser shelf water. River plumes can be categorized into two major types: surface-advected and bottom-trapped plumes (Yankovsky & Chapman, 1997, hereinafter noted as YC97). For a surface-advected plume, it tends to spread radially from its source and extend well offshore, typically forming an anticyclonically rotating bulge that is attached to the river mouth (Chao & Boicourt, 1986; Fong & Geyer, 2002; Horner-Devine et al., 2006). In this case, the plume water floats over denser ambient seawater as a thin layer and has little contact with the bottom topography. For a bottom-trapped plume, as the main part of the plume contacts the sloping bottom, a density front extends from the surface to the bottom with a strong horizontal gradient, and the trapping depth of plume exceeds the inflow depth. Many large midlatitude
rivers that discharge into a mildly sloping shelf feature bottom-trapped plumes, including the Mississippi River plume (Cochrane & Kelly, 1986; Murray, 1998; Z. Zhang & Hetland, 2012; X. Zhang et al., 2012), the Niagara River plume (Horner-Devine et al., 2008; Masse & Murthy, 1992), and the Rhine ROFI (region of freshwater influence; De Boer et al., 2009; Simpson et al., 1993). Terrestrial materials, including sediments, nutrients, and organic matter, are transported in both along-shelf and cross-shelf directions by means of advection and mixing in bottom-trapped plumes as well as their fronts. Those terrestrial materials therefore have a profound influence on shelf circulation, coastal ecosystem health, and seafloor morphology, contributing to harmful algae blooms, hypoxia events (Bianchi et al., 2010; D. Li et al., 2002; Rabouille et al., 2008; Zhou et al., 2008), and sediment deposition on the continental shelf (Geyer et al., 2004; Milliman et al., 1985; Wright & Nittrouer, 1995).

In comparison with surface-advected plumes, bottom-trapped plumes are relatively stable and can impose persistent impacts on coastal areas of continental shelves. The steadiness of a bottom-trapped plume is due to the trapping of its plume front. The down-shelf current in the bottom boundary layer (BBL) of a bottom-trapped plume front induces an offshore Ekman transport, which pushes the plume front seaward. When the bottom front reaches to a certain depth (h₀), the down-shelf currents in the BBL of the plume front relaxes or reverses, which shuts down the offshore Ekman transport. Therefore, the plume front is trapped along the isobath of trapping depth h₀ (Chapman & Lentz, 1994). Based on the balance of frontal transport and the outflow at the river mouth, YC97 provided a theoretical estimation for the trapping depth; that is, \( h₀ = \sqrt{2TF/g} \), where \( T \) is the outflow transport, \( f \) is the Coriolis parameter, and \( g' \) is the reduced gravity of the river outflow. The trapping depth signifies the cross-shelf scale of a bottom-trapped plume and is also associated with many local features such as convergence, upwelling, and surface-intensified current (Chapman & Lentz, 1994).

According to the trapping mechanism, the trapping depth is determined by its buoyancy inflow, latitude, and plume density. However, other external forces can also modulate a bottom-trapped plume front. Upwelling-favorable winds tend to spread the plume offshore and weaken or reverse its down-shelf propagation, while downwelling-favorable winds constrain the plume in coastal areas and augment its down-shelf propagation (Fong & Geyer, 2001; Pimenta & Kirwan, 2014; Whitney & Garvine, 2005). By introducing the wind-driven entrainment, Lentz (2004) suggests that entrainment is also a significant process that spreads plume water offshore under upwelling-favorable winds. Model results in Chao (1990) show that tidal-induced residual eddies can enhance plume bulge. Isobe (2005) points out that along-shelf tidal current helps suppress inertial instability in the bulge region, thus increasing the proportion of plume water that extends down-shelf. The tidal effect of stabilizing the plume bulge and augmenting the down-shelf buoyant coastal current is significant in the Changjiang River plume (M. Li & Rong, 2012). S. N. Chen (2014) suggests that the along-shelf tidal current can also enhance down-shelf coastal current by increasing the incident angle of the bulge current. Similar studies on tidal modulation can be found in the Chesapeake Bay outflow plume (Guo & Valle-Levinson, 2007) and Rhine ROFI (Simpson et al., 1993). Although there is little doubt that tidal currents can augment the down-shelf buoyant coastal current, the majority of previous studies focus on bulge regions (e.g., Chao, 1990; S. N. Chen, 2014; Isobe, 2005; M. Li & Rong, 2012).

Numerical studies frequently reported that, when external forces (such as wind or tide) were removed, a great portion of the simulated river plume propagated up-shelf continuously, which was uncommon in the observations (Chapman & Lentz, 1994; Garvine, 1999, 2001; Matano & Palma, 2010; Wu et al., 2011; Yankovsky, 2000). Such unrealistic up-shelf extension reduced the down-shelf freshwater flux of the buoyant coastal current (Garvine, 2001). Therefore, in many numerical studies, it was a common practice to add an external down-shelf mean flow solely to arrest the up-shelf intrusion of a river plume and to get a well-developed bottom-trapped plume (e.g., Fong & Geyer, 2002; Narayanan & Garvine, 2002; Pimenta & Kirwan, 2014; Yankovsky & Chapman, 1997). Although bottom-trapped plumes are widely observed, it is difficult for idealized simulations to reproduce them without external forces. That is to say, not only the river itself but also some other external processes should be included to form a down-shelf propagating bottom-trapped plume.

In this study, the Changjiang River plume was used as an example to determine the key processes that maintain the bottom-trapped plume. The Changjiang River plume is a highly dynamic system. Since the 1960s when Mao et al. (1963) first depicted the expansion patterns of the plume, extensive efforts have been devoted to studying how the surface part of the plume expends over the Yellow Sea and the East China
Sea. So far, the plume is believed to extend in three major pathways: southward along the Zhejiang-Fujian (also known as Zhe-Min) Coastal Water (ZMCW) mainly in nonsummer seasons that forms the Zhe-Min Coastal Current (ZMCC), northeastward toward Jeju Island in summer (Beardsley et al., 1985; Chang & Isobe, 2003; Limeburner et al., 1983; Mao et al., 1963; Zhao, 1991; J. Zhu et al., 1997), and northward along the Jiangsu Coast in multiple seasons (Wu et al., 2014). It was conventionally believed that the ZMCC prevails in winter but vanishes in summer, as controlled by the Eastern Asian Monsoon. However, recent studies show that buoyancy plays an essential role in the water transport of the ZMCC (Wu et al., 2013). Occasionally a down-shelf buoyant coastal current can be detected in the ZMCW even in summer. P. Li et al. (2014) analyzed the acoustic Doppler current profiler (ADCP) data in July and August at three sites in the ZMCW and found that the residual currents were stably southward at two sites (Figure 1a).

It is unclear how a bottom-trapped plume can persist in such an energetic environment, as there is no steady, down-shelf background flow and the ambient current, that is, the Taiwan Warm Current, is actually up-shelf. The wind forcing has dramatic seasonal variations, switching between downwelling-favorable winds in winter and upwelling-favorable winds in summer. Even the buoyancy inflow has evident seasonal variations. The discharge of Changjiang River ranges from ~1 × 10^4 m^3/s in the dry winter season to ~5 × 10^4 m^3/s in the summer flood season. Therefore, the mechanisms maintaining the bottom-trapped plume remain unclear.

To address this issue, this study focused on the strength and trapping depth of the buoyant ZMCC. A numerical model was used to diagnose the evolution of the ZMCC as functions of the Changjiang River discharge, East Asia Monsoon, and tidal forcing, in addition to exploring the mechanism that modulates the plume front and its trapping depth. The remainder of this paper is structured as follows. Settings and validations of numerical experiments are introduced in section 2. Section 3 describes basic facts of the Changjiang River plume in the ZMCW and its spatial-temporal variations. Driving forces of the ZMCC and mechanisms that sustain the bottom-trapped plume are discussed in section 4. Finally, conclusions are drawn in section 5.

2. Materials and Methods

2.1. Study Area

This study focuses on the shelf water shallower than 100 m in the East China Sea (Figure 1a), that is, the ZMCW. The extent of the Zhe-Min Coast is ~500 km. Isobaths shallower than 50 m are in line with the coast south of the Zhoushan Islands. The ZMCW is highly influenced by the Changjiang River plume and the Taiwan Warm Current. Because of the huge amount of terrestrial materials transported by the plume, the ZMCW suffers from frequent harmful algae blooms and seasonal hypoxia (Figure 1b; Cai et al., 2011; Chai et al., 2006; D. Li et al., 2002; Rabouille et al., 2008; Zhou et al., 2008; J. Zhu et al., 2016). There is a long mud belt lying on the ZMWW, which was formed as a result of sediment deposition from the Changjiang River (Figure 1b; Liu et al., 2018, 2006; Milliman et al., 1985; Xu et al., 2009; C. Zhu et al., 2011).

The plume propagates down-shelf along the Zhe-Min Coast in autumn, winter, and early spring, forming the ZMCC (Beardsley et al., 1985; Milliman et al., 1985; Niino & Emery, 1961; Wu et al., 2013). The Taiwan Warm Current, which originates from the Taiwan Strait and the Kuroshio Branch Current north of Taiwan, flows northeastward parallel to the 50-m isobath in the offshore region of the ZMWW (J. Zhu et al., 2004). Where the fresher and lighter plume water meets the saltier and denser ambient water, strong salinity and density fronts form in the ZMWW.

There is a considerable tidal forcing in this area, and the direction of the tidal current is perpendicular to the coast and isobaths (Figure 1b). The tidal current speed in this area reaches O(1) m/s (Wu et al., 2013). The tide type is primarily regular semidiurnal with a tidal form factor $A_{K1}/A_{M2}+A_{S2}$ less than 0.4 (Su & Yuan, 2005). The East Asia monsoon prevails in the ZMWW, which includes the southerly summer monsoon and the northerly winter monsoon (in nonsummer seasons).

2.2. Numerical Model

The numerical model used here was originated from ECOM-si that was developed by Blumberg (1994). It has been reconstructed and improved by the State Key Laboratory of Estuarine and Coastal Research (SKLEC), East China Normal University (Wu & Zhu, 2010). Specifically, a third-order advection scheme HSIMT (Wu & Zhu, 2010) was developed to solve the tracer equations. The Mellor-Yamada 2.5 turbulence closure model (Galperin et al., 1988; Mellor & Yamada, 1982) and the Smagorinsky scheme (Smagorinsky, 1963) were
used to calculate vertical and horizontal mixings, respectively. A Yellow and East China Seas model was con-
figured (Wu et al., 2011), which covered the entire East China Sea, Yellow Sea, and Bohai Sea, as well as part of
the Japan Sea and the Pacific Ocean. The model resolution was 1.5 km or higher around the Changjiang River
Estuary and was 3–4 km in the ZMCW (Figure 1c). The 20 $\sigma$ layers were set in the vertical direction with re-
fined surface layers. The hydrodynamic open ocean boundary was driven by the momentum flux, which contained
both the tidal current and shelf circulations. The heat flux was calculated with the bulk formula suggested by
Ahsan and Blumberg (1999), based on the modeled sea surface temperature and atmospheric parameters
(mean sea level pressure, 2-m temperature, 2-m relative humidity, and 10-m wind and total cloud cover).
The open boundary conditions of shelf currents, salinity, temperature, and the initial conditions of salinity
and temperature were derived from Simple Ocean Data Assimilation. The initial conditions of currents and
elevation were simply set as 0, because the adjustment of hydrodynamic processes was rapid. For more
details of the model please refer to Wu et al. (2011, 2014).

2.3. Model Validation
The tidal system and shelf circulation have been well validated by Wu et al. (2011). The simulated tidal eleva-
tions at 10 stations around the Changjiang River Estuary and the Zhe-Min Coast matched in situ data well (see
Table 2 in Wu et al., 2011). The realistic and climatological near-surface salinity distribution was also well
reproduced by the model with acceptable accuracy in each season (Wu et al., 2014).
However, those previous validations mainly focused on near-surface plume features. Since this study focused on the bottom-trapped plume, the model was further validated with bottom salinity derived from conductivity-temperature-depth (CTD) data collected during eight survey cruises from 2012 to 2016. These surveys were conducted in the months of February, March, May, July, and August and were operated by the National Natural Science Foundation of China and the SKLEC. The sampling sites from these surveys are shown in Figure 2a, which covered the area influenced by the Changjiang River plume in the Jiangsu, Shanghai and Zhejiang coastal waters. The model ran from 1 January 2011 for 6 years, driven by the realistic Changjiang River discharge measured at the Datong Station (operated by the Changjiang Water Resource Commission) and the ERA Interim reanalyzed atmospheric data from the European Centre for Medium-Range Weather Forecasts.

The modeled surface and bottom salinity data were compared with in situ data. To better visualize the results, normalized standard deviation, correlation coefficient (CC), and normalized root-mean-square error (NRMSE) were calculated and shown as a Taylor diagram (Taylor, 2001; Figure 2b). As is revealed in the Taylor diagram, for bottom salinity (opaque symbols in Figure 2b), CCs were above 0.8 and the NRMSEs were below 0.8 in all cases, indicating that the model captured the bottom salinity patterns quite well. For the surface salinity (semi transparent symbols in Figure 2b), CCs were not as high as for bottom salinity, since the surface river plume was much more variable. Nevertheless, for all cases, CCs were higher than 0.75. NRMSEs of surface salinity were under 0.9. Therefore, the surface salinity pattern was also well reproduced by the model.

The bottom and surface salinities from all eight surveys were pooled together and compared with the in situ data (Figures 2c and 2d). The CCs, root-mean-square errors, and skill scores (SSs; Murphy, 1988) between modeled and in situ data were calculated. Performance levels were graded by SS as follows: >0.65 (excellent); 0.65–0.5 (very good); 0.5–0.2 (good); <0.2 (poor; Allen et al., 2007; Maréchal, 2004; Ralston et al., 2010). SSs were above 0.75 for both bottom and surface layers, which means that the model can reliably reproduce the Changjiang River plume. The root-mean-square errors and CCs for both surface and bottom salinity were also reasonable. Regression lines of the scatter points have slopes close to one (Figures 2c and 2d), which means no systematic errors were made by the model.

Figures 2e and 2f show the validation results of the model against the salinity data collected by SKLEC cruise in 6–25 May 2016 (yellow diamonds in Figures 2a and 2b). Overall, the model well reproduced the extent, location, and pattern of the bottom salinity front outside the Changjiang River Estuary and in the ZMCW. The salinity was slightly underestimated north of 32°N, which was outside the focus area of the study. From Figures 2e and 2f, it can be seen that there was a continuous bottom salinity front along the bathymetry, implying that the plume was in a bottom-trapped status. The better performance of the numerical model for bottom salinity also implied that the salinity front was less variable at the bottom than at the surface.

### 2.4. Numerical Experiment Settings

A control run and several idealized experiments were conducted. The control run (Exp0) was driven by climatological monthly mean Changjiang River discharge (based on historical observations in the Datong Station) and climatological monthly mean atmospheric data (from European Centre for Medium-Range Weather Forecasts). Exp0 was spun up for one year, and the results in the second year were used for analysis. Several idealized experiments under different tidal, wind, and runoff conditions were set up for diagnosing plume and transport dynamics in the ZMCW. The settings of these experiments are listed in Table 1 and will be further discussed in a later section.

### 2.5. Frontal Detection

The gradient method was used to detect the modeled bottom plume front, which was successfully applied in recognizing oceanic fronts (He et al., 2016; Huang et al., 2010; Tang, 1995; Tang & Zheng, 1990; Wall et al., 2008). First, the modeled salinity was interpolated into regular grids (0.02° × 0.02°) in the latitude-longitude coordinate. Then, the salinity gradient magnitudes (SGMs) were calculated for all 3 × 3 grids using equation (1):

\[
SGM = \sqrt{(\Delta S_1/\Delta x_1)^2 + (\Delta S_2/\Delta x_2)^2}
\] (1)
Figure 2. Model validation using in situ conductivity-temperature-depth (CTD) salinity data from eight survey cruises. Dots in (a) indicate the coverage and sampling sites of the eight surveys. The Taylor diagram (b) shows model performance on surface (semitransparent shapes) and bottom (opaque shapes) salinity in each survey. Scatters of model versus in situ surface and bottom salinity are plotted in (c) and (d). Red and blue solid lines are the regression lines of scatter points in each plot, and the regression equations are shown below the scatter points. In situ and modeled bottom salinities in the May 2016 survey are shown in (e) and (f). SS = skill score; RMSE = root-mean-square error.
where $\Delta x_1$ and $\Delta x_2$ indicate the two diagonal distances between the corner points of a 3 $\times$ 3 grid and $\Delta S_1$, $\Delta S_2$ are the diagonal salinity differences, respectively.

As is suggested by Fedorov (1986), the grid point can be considered as being in the frontal zone when the SGM exceeds 10 times of the average gradient (Fedorov threshold) in the study area. The occurrence probability of the front (OPF; He et al., 2016) was defined as the percentage occurrence of SGM being greater than the Fedorov threshold at each grid in the simulated period. We used the OPF to represent the strength of bottom salinity fronts rather than the averaged SGM, since the later would smooth and widen the salinity fronts.

### 3. Results

#### 3.1. The Bottom River Plume Front in the ZMCW

The bottom OPF for each season in the ZMCW was calculated based on the modeled bottom salinity from Exp0 (Figure 3). Spring was defined as March to May, summer as June to August, autumn as September to November, and winter as December to the following February. The solid black contour lines denote a 50% occurrence probability, which was considered as the threshold of strong frontal zones. Regions with OPF of 0–50% were considered as weak frontal zones.

The bottom front was trapped between the 20- and 40-m isobaths in all four seasons. The shape of the bottom frontal zone was similar to that observed in May 2016 (Figure 2e); however, the surface plume (shown by 30-psu isohalines) was much more variable. There exists a bottom front between 29°N and 30°N all year round, even in summer when the plume mainly spreads in a northeastward direction offshore (Beardsley, 1983; Kim et al., 2009; Lie et al., 2003; Limeburner, 1983; Mao et al., 1963; Wu et al., 2014). The bottom front reached 26°N from autumn to the following spring. The strength and width of the bottom front reached its peak in autumn. In the following winter and spring, the extension, strength, and width of the bottom front decreased slightly. In summer, the southern end of the bottom front shrunk to 28°N.

The bottom front is an important characteristic for a plume, which indicates how far the main part of a bottom-trapped plume can reach in an along-shelf direction. It also indicates the center of the associated buoyant coastal current. As suggested by the bottom front in the ZMCW, the Zhe-Min branch of the Changjiang River plume existed in all seasons, even in summer when the wind was upwelling favorable. The Zhe-Min plume branch prevailed in autumn and was gradually weakened in the following seasons, and then finally shrunk to its smallest scale in the summer. No matter in which direction the surface plume expanded, the bottom plume in the ZMCW was always trapped within the 40-m isobath. The persistence of the bottom front indicated that the ZMCC was present in all seasons.

#### 3.2. Cross-Sectional Plume Structure in the ZMCW

Figure 4 shows the subtidal salinity and current profiles in Section A (delineated in Figure 1a) in spring and neap tides during winter (February) and summer (August). The section was chosen to be perpendicular
with the isobaths and the frontal zone. The tidal-averaged salinity front was constantly present in Section A. The tidal bottom front was ~20 km wide and was trapped between the 20- and 40-m isobaths, whereas the location of surface plume front varied with seasons (Figure 4). In winter, the surface plume front was ~5 km further offshore than the bottom front. However, in summer, the distance between the surface and bottom plume fronts rose to ~40 km. The baroclinic Rossby deformation radius, $R_d = \sqrt{g h f}$, was ~14 km. Ideally, the width between surface and bottom fronts, noted as $W$, should be similar to $R_d$. However, $W$ was much smaller than $R_d$ in winter and much larger in summer. This suggested that different winds in winter (downwelling favorable) and summer (upwelling favorable) significantly change the location of surface plume front. On the contrary, the location of the bottom plume front was less affected.

The along-shelf current, which was tidal averaged to remove the tidal signals, was surface intensified in the frontal zone. A secondary intensification was also observed on the top of BBL, especially during the spring tide in summer (Figure 4d), which was consistent with the findings in P. Li et al. (2014). The surface part of the along-shelf frontal current was down-shelf with speeds exceeding 0.4 m/s in both seasons.
During neap tides, the along-shelf frontal current reversed to the up-shelf direction in the BBL of the front, which is in agreement with the bottom-trapped plume theory described in Chapman and Lentz (1994). However, during spring tides, the along-shelf residual current remained down-shelf in the onshore half of the frontal BBL. In addition, there was upwelling in the seaward edge of the bottom front and downwelling in the shoreward edge of the bottom front. According to Chapman and Lentz (1994), the upwelling and downwelling in the frontal zone resulted from the convergence and divergence of Ekman transport in the BBL during the trapping of the plume front.

In winter (Figures 4a and 4c), the isohalines in the plume front were nearly vertical, showing characteristics of a bottom-trapped plume front under the influence of downwelling-favorable winds. In summer (Figures 4b and 4d), the trapping depth of the plume front was slightly deeper than that in winter and the slope of the front also decreased. The water column became strongly stratified offshore of the bottom front. In relative terms, the trapping depth of the plume front was deeper and the associated salinity gradient was weaker during spring tides than during neap tides (Figures 5). During spring tides, the isohalines in the BBL became vertical due to strong mixing. However, above the BBL, the slope of the front hardly changed under different tidal amplitudes.

To diagnose the dynamic mechanisms, we outputted the momentum terms from the model simulation:

\[
\frac{\partial \mathbf{v}}{\partial t} + (\mathbf{v} \cdot \nabla) \mathbf{v} = -g\nabla \eta - \frac{g}{\rho_0} \int \nabla \rho dz + \mathbf{F}_r - 2\Omega \times \mathbf{v} \tag{2}
\]

in which \(\frac{\partial \mathbf{v}}{\partial t}\) is local acceleration, \((\mathbf{v} \cdot \nabla)\mathbf{v}\) is advection, \(-g\nabla \eta\) is barotropic pressure gradient force, \(-\frac{g}{\rho_0} \int \nabla \rho dz\) is baroclinic pressure gradient force, \(-2\Omega \times \mathbf{v}\) is Coriolis force, and \(\mathbf{F}_r\) is friction. The cross-shelf component of momentum terms in Section A was tidal averaged during
the same periods as in Figure 4, shown as Figures 6 and 7. The tidal-averaged local acceleration and horizontal friction terms were negligible.

The two terms composing thermal wind balance were plotted to examine how close the velocity shear in the plume front follows thermal wind shear (Figure 8):

\[ \frac{\partial v}{\partial z} = \frac{g}{\rho_0 f} \frac{\partial \rho}{\partial x} \]  

(3)

**Figure 6.** Tidal mean cross-shelf profiles of momentum terms in Section A in August (a–e) and February (f–j) during neap tide from Exp0. (a, f), momentum advection, (b, g) barotropic gradient force, (c, h) baroclinic gradient force, (d, i) Coriolis force, and (e, j) vertical friction. Black contours indicate salinity profiles.
where $\frac{\partial}{\partial z}$ is vertical velocity shear and $-\frac{F}{\rho_0 g}$ represents thermal wind shear. These two terms were calculated on each point in the frontal region in Section A during the same time periods as the salinity-velocity profile (Figure 4).

The plume was basically in geostrophic balance during neap tides (Figure 5). The momentum was mainly balanced by barotropic pressure gradient force, baroclinic pressure gradient force, and Coriolis force, in agreement with the momentum balance of a theoretical bottom-trapped plume front (Chapman & Lentz, 1994). The seaward barotropic pressure gradient force, cooperating with the Coriolis force,
induced a down-shelf current in the plume front. Because of the bottom-intensified shoreward baroclinic gradient force in the frontal zone, the down-shelf current became surface intensified. The core of the down-shelf current was located above or near the seaward edge of the bottom front, where the cross-shelf density gradient was strongest. Above the BBL (i.e., large height above bottom), where the core of this buoyant coastal current is located, the vertical velocity shear basically followed the thermal wind shear (Figure 8). However, the thermal wind balance was broken within the BBL, which was more evident during the spring tide (Figures 8c and 8d), due to the enhanced momentum advection and bottom friction (Figure 7). This indicated that the strong tidal forcing was important in modulating the momentum balance, which was consistent with the previous findings in the Norfolk Sandbanks (Huthnance, 1973) and Georges Bank (C. Chen & Beardsley, 1995; Loder, 1980). In the barotropic condition, Loder (1980) pointed out that the cross-isobath tidal current induces a subtidal momentum flux convergence due to continuity, which drives a down-shelf residual current under the Coriolis forcing. This process was termed as tidal rectification, which is also regulated by bottom friction. When the vertical stratification is included, the tidal rectification will generate a subsurface intensified down-shelf residual current on the top of the BBL over the slope (C. Chen & Beardsley, 1995). Such a phenomenon was also observed in this study (Figure 4d). As a result, the down-shelf current is intensified not only at the surface due to the thermal wind balance but also on the top of the BBL due to tidal rectification (Figure 4d). Nevertheless, the momentum advection was much smaller than the baroclinic pressure gradient force in winter (Figures 7f and 7h). Although the tidal rectification induced a down-shelf current, the core of the ZMCC was still under geostrophic balance.
3.3. Variations of the Bottom Plume Front in the ZMCW

Empirical orthogonal function (EOF) was used to further analyze variations in the bottom front gradient in the ZMCW and thereby to find clues for revealing the formation and modulation mechanisms of the bottom front in this area. Using daily subtidal bottom salinity fields from Exp0, the subtidal bottom salinity gradient fields of the ZMCW were calculated using the method described in section 2.5. The annually averaged salinity gradient was removed from the salinity gradient field before the EOF calculation.

The standard deviation of bottom salinity gradients is shown in Figure 9a. Generally, the bottom salinity gradient of the frontal zone had a higher standard deviation. Although the bottom front was robustly trapped between the 20- and 40-m isobaths, its strength, extent, and location varied with time. The three leading EOF modes of the bottom salinity gradient anomaly are shown in Figures 9b–9d. The time series of the three EOF modes are shown in Figures 9e–9g. Their contributions to the total variance were 40.2%, 24.9%, and 10.1%, respectively, accounting for 75.2% of the total variance altogether. The Changjiang River discharge, wind forcing, and tidal forcing were found to be the three main factors contributing to those variations.

Mode 1 shows the summer-winter cycle of the bottom front in the ZMCW. During spring and summer seasons (i.e., from May to early September), the southward extent of the bottom plume front was much smaller than during the autumn and winter seasons. In the latter two seasons, the front prevailed all along the coast. This pattern was generally in coherence with the annual variation of the East Asia Monsoon (see vectors in Figure 9e). Down-shelf plume propagation was weakened in the summer, due to the upwelling-favorable summer monsoon. While in winter, downwelling-favorable winds pushed the plume shoreward and promoted down-shelf propagation. However, it is notable that most of the variation occurred south of 28°N. North of this point the variation is not evident, suggesting that the plume was persistently bottom-trapped in that area.

Mode 2 represents the sudden growth of the bottom front in September and October. The front reached its peak in the middle of October. This pattern was driven by both high levels of Changjiang River discharge and downwelling-favorable winds (see vectors in Figure 9e and the blue curve in Figure 9f). In September, the
wind turned northerly and pushed the remaining massive summertime river plume to propagate down-shelf. In addition, the Changjiang River discharge was still high in September and October. Owing to the combined effects of high buoyancy supply and downwelling-favorable winds, the bottom front rapidly grew and reached its annual peak in September and October. In the following months, despite the strong northerly wind, the bottom front gradually declined due to the reduced Changjiang River discharge.

Mode 3 indicates the cross-shelf movement of the bottom front in the ZMCW. The time series of this mode is dominated by a strong spring-neap tidal signal. That is to say, the cross-shelf movement of the bottom plume front was highly correlated with the tidal forcing. During spring tides, the bottom front in the ZMCW tended to move seaward, while the trapping depth decreased during neap tides. Besides having a spring-neap cycle, the cross-shelf movement of the bottom front also featured a distinct solstitial-equinoctial variation. The bottom front was more offshore in late March (vernal equinox) and late September (autumnal equinox), when tidal forcing reached its annual peaks.

Buoyancy inflow and wind forcing are believed to have strong influences on the trapping depth of a bottom-trapped plume front (Fong & Geyer, 2001; Pimenta & Kirwan, 2014; Whitney & Garvine, 2005; Yankovsky & Chapman, 1997). Although the Changjiang River discharge has a large annual variance (~40,000 m$^3$/s) and wind forcing also varies dramatically, the trapping depth of the bottom front in the ZMCW shows little seasonal variation. However, as is suggested by Mode 3, the plume’s trapping depth seems to be controlled by tidal forcing. More in-depth discussions on the formation and sustention mechanisms of the bottom-trapped Changjiang River plume in the ZMCW will be provided in the next section.

4. Discussions

4.1. Responses of the Plume and Associate Coastal Current to Runoff and Wind Forcings

The above results indicated that the bottom plume front was much more stable than the surface front. Nevertheless, the trapping depth, intensity, and extent of the bottom plume front still had seasonal and fortnightly variations, as shown in the EOF results. The first two EOF modes, which account for more than 65% of the bottom front variations, are due to seasonal changes in the Changjiang River discharge and wind forcing. To understand how the Changjiang River discharge and wind forcing influenced the bottom front, two sets of numerical experiments (Exp1-Exp2 and Exp3-Exp4, see Table 1) were set up. In addition, a baseline experiment, that is, Exp5, was set, where wind forcing was excluded and the Changjiang River discharge was set to a medium value (3 × 10$^4$ m$^3$/s). Tidal forcing was included in all five experiments.

In Exp1 and Exp2, the Changjiang River discharges were set to a dry season value (1 × 10$^4$ m$^3$/s) and a flood season value (5 × 10$^4$ m$^3$/s), respectively. Wind forcing was removed from both experiments. As is revealed in OPF results of Exp1, Exp5, and Exp2 (Figures 10a–10c), the strength and extent of the bottom front in the ZMCW increased along with the Changjiang River discharge. At the same time, the surface plume also expanded dramatically. Hence, the Changjiang River discharge determined the overall scale of the plume from near field to far field. In Exp3 and Exp4, wind forcing was set to summer monsoon conditions (upwelling-favorable southerly wind) and winter monsoon conditions (downwelling-favorable northerly wind), respectively. The Changjiang River discharge was set to a medium value (3 × 10$^4$ m$^3$/s). As is shown in OPF results of Exp3, Exp5, and Exp4 (Figures 10d–10f), wind forcing strongly impacted the distribution of the plume. Upwelling-favorable summer wind shortened the along-shelf scale of the plume and spread the surface plume offshore, while downwelling-favorable winter wind confined the plume within the coastal area and elongated its along-shelf scale, which is in keeping with previous theoretical studies (e.g., Simpson, 1997; Lentz, 2004; Whitney & Garvine, 2005; Pimenta & Kirwan, 2014).

The ZMCC was previously considered to be a wind-induced coastal current under the northerly winter monsoon. However, it has been reported that buoyancy plays an essential role in driving the southward coastal current (Wu et al., 2013). In addition, the momentum analysis described above gave us the idea that the ZMCC is under the control of barotropic and baroclinic pressure gradient forces, rather than wind stress. To obtain a better picture of how the ZMCC evolves and the underlying dynamics, the subtidal surface and bottom salinities, bottom salinity gradient, and along-shelf residual volume flux in Section A from Exp0 are plotted in Figures 11a–11c and 11e–11g. The along-shelf residual volume flux was calculated as follows:
Residual volume flux \( \eta \) is free surface height, \( h \) is tidal mean water depth, \( V \) is the along-shelf component of horizontal velocity (negative values indicate down-shelf velocity), and \( \langle \rangle \) is a low-pass filter operator with a cutoff window of 75 hr. An idealized experiment, Exp6, was set to simulate a purely wind-driven current in the ZMCW. In Exp6, Changjiang River runoff was turned off and the initial salinity and temperature were set as constant to exclude the buoyancy effect. The down-shelf volume transports in Exp0 and Exp6 are plotted in Figure 11i, together with the wind vector and Changjiang River discharge. Generally, in Exp0, the bottom front in the ZMCW was located between the 20- and 40-m isobaths. The core of the ZMCC was located around the 30-m isobaths, above the bottom front (Figures 11b, 11f, 11c, and 11g). The down-shelf transport persisted throughout four seasons except for a few interruptions in late spring and summer (Figures 11g and 11i). The mean down-shelf volume transport of the ZMCC across Section A in winter was \( \sim 0.2 \) Sv (Figure 11i), which was consistent with survey results from the same area obtained by Wu et al. (2013). The bottom front reached its peak in autumn and early winter (September to December), and the ZMCC reached maximum transport in the meantime. The direct wind-driven transport only covered less than 23% of the annual transport of the ZMCC (Figure 11i). The down-shelf wind-driven current started in the middle of August when the summer monsoon abated, prevailed through autumn and winter when the northerly winter monsoon was dominant, and vanished in spring and summer when it shifted to the southerly summer monsoon. In autumn and early winter, when the ZMCC reached its peak, the direct contribution of wind-driven transport to the total down-shelf transport was only \( \sim 15\% \). In spring and summer, the wind-
driven contribution even decreased to 0 (Figure 11i). In brief, direct wind-driven transport only accounted for a very small fraction of the ZMCC.

Nevertheless, as is shown in Figures 10d–10f, wind forcing had a great influence on modulating the buoyancy supply to the ZMCW. In spring and summer when the winter monsoon ceased and the summer monsoon prevailed, wind forcing gradually turned upwelling favorable. The upwelling-favorable wind drove the majority of the river plume offshore, which decreased the amount of buoyancy transported to the ZMCW and thus reduced the intensity of the plume front there. Therefore, due to the decline of buoyancy supply, the ZMCC became weaker in spring and summer. While in autumn, the wind direction turned northerly, and the offshore plume water was pushed back to the coastal area by shoreward wind-driven Ekman transport. Increased buoyancy caused a stronger plume front in the ZMCW, allowing the ZMCC to reach its annual peak in autumn. In winter and spring, the buoyancy moved further down-shelf and the Changjiang River discharge decreased. Hence, the plume front in the ZMCW was weakened and the ZMCC slowed down. In summary, the ZMCC was directly driven by buoyancy, but buoyancy itself was modulated by wind forcing and Changjiang River discharge.

4.2. Sustention of the Plume Front and the ZMCC

Summer counter-wind ZMCC was reproduced by the model (Figures 11g and 11i), which was consistent with the observation reported by P. Li et al. (2014; also see Figure 1a). There are various ways of quantifying the relative importance of wind forcing and buoyancy forcing, such as Wedderburn number in enclosed lakes (Shintani et al., 2010; Thompson & Imberger, 1980; Wedderburn, 1912) and the Whitney and Garvine (2005) parameter \( W_r \) for buoyant coastal currents. Whitney and Garvine (2005) suggested that the upwelling-favorable wind reverses the buoyant coastal current only when the wind is strong enough so that \( W_r = \frac{\mathbf{u}_{\text{wind}}}{\mathbf{u}_{\text{buoyancy}}} \) exceeds 1. Here \( \mathbf{u}_{\text{wind}} \) is the wind-driven along-shelf velocity that can be calculated with an empirical formula, and \( \mathbf{u}_{\text{buoyancy}} \) is the buoyancy-driven velocity that can be calculated with the thermal-
Figure 12. Comparison of predicted trapping depth calculated with YC97 (blue line) and modeled trapping depth where the maximum bottom salinity gradient took place (red line) in Section A from Exp0. Green dashed lines indicate spring tides. When calculating the theoretical trapping depth, T accounts for down-shelf transport cross section A and $g'$ is calculated using the local plume density anomaly of Section A.

wind depth, local down-shelf transport ($T$) and local reduced gravity ($g'$) were used instead of the outflow and reduced gravity from the river mouth. This was because YC97 assumes that the flux and density anomaly across the down-shelf cross section equals those from the river mouth, which is not realistic. Although YC97 does not include tidal forcing, there were still spring-neap variations in predicted trapping depth, due to spring-neap variations in local transport and reduced gravity under the tidal effects.

There were large discrepancies between the theoretical and modeled trapping depth, especially in spring and summer (Figure 12). The theoretical trapping depth was $<10$ m in late spring and summer, shallower than the river mouth depth, which means that the transport could no longer support a bottom-trapped river plume, and the plume should be transformed into a surface-advected one. However, the modeled trapping depth could still persist $\sim 25$ m and the plume remained bottom trapped. Based on the geostrophic balance, ZMCC persists as long as the bottom density gradient occurred. When calculating the theoretical trapping depth, local down-shelf transport ($T$) and local reduced gravity ($g'$) were used instead of the outflow and reduced gravity from the river mouth. This was because YC97 assumes that the flux and density anomaly across the down-shelf cross section equals those from the river mouth, which is not realistic. Although YC97 does not include tidal forcing, there were still spring-neap variations in predicted trapping depth, due to spring-neap variations in local transport and reduced gravity under the tidal effects.

To determine the mechanism sustaining the bottom-trapped plume in the ZMCC, we look back into the EOF results of the bottom salinity gradient (Figure 9). The EOF results indicated that there were no significant differences in trapping depth between seasons regardless of changeable wind and runoff, but there were significant variations on a fortnightly basis (Figure 9). Spring-neap and solstitial-equinoctial variations were both evident in the modeled trapping depth (Figure 12). Therefore, tidal forcing should play an important role in controlling the trapping depth. The spring-neap variations of bottom fronts can be found in many tide-induced thermal fronts (e.g., Horsburgh & Hill, 2003; Luyten et al., 2003; Lwiza et al., 1991; Yu et al., 2016). A basic mechanism for the fortnightly moving of the bottom tidal front is that the coastal mixing areas are enlarged in spring tides due to the thickened BBL. The location of the tidal front can be predicted by the Simpson-Hunter (SH) number ($SH = \log(h/u^*)$, where $h$ is tidal mean water depth and $u$ is the tidal current amplitude), which indicates whether tidal mixing can overcome the potential energy in a water column and mix it into a vertical homogeneous state (Simpson & Hunter, 1974). Figure 13 shows the SH number in the Yellow and East China Seas. Highlighted areas between dashed and solid black lines indicate a predicted tidal-induced mixing front locus ($SH = 1.7-2.0$) during spring tides. It is marked that the predicted tidal-induced front locus in the ZMCC is in line with the bottom plume front. Plume water is also a great source of potential energy. Therefore, as long as plume water reaches the ZMCC, a tidal-induced plume front will be constructed at the predicted tidal-induced front locus.

To further understand how tidal forcing regulated the trapping depth of the plume in the ZMCC, the BBL thickness (determined following Weatherly & Martin, 1978, the height at which the time-averaged turbulent kinetic energy goes to 0), the cross-shelf volume flux in the BBL, and the salinity gradient in Section A from Exp5 during both spring and neap tides are plotted in Figure 14. The 1:1 straight line in Figure 14b means that BBL thickness equaled water depth, indicating complete mixing of the water column. During spring tide, the area with complete mixing (i.e., shoreward of 23 m) was much larger than that during...
neap tide (i.e., shoreward of 17 m). Seaward of the complete mixing area, the BBL thickness decreased rapidly and then reached a relatively stable value, while slightly increasing offshore. The BBL was approximately twice thicker during the spring tide than during the neap tide (Figure 14b). The spring-neap variation of BBL thickness was also evident in Exp0 (see vertical eddy viscosity profile in Section A from Exp0 in Figure 15).

According to the frontal trapping theory (Chapman & Lentz, 1994), a bottom-trapped plume front is moving offshore initially under the bottom Ekman transport. During this process, up-shelf bottom counter flow occurred in the seaward portion of the front, which induced an onshore Ekman transport. Finally, the front is trapped at a certain isobath where the onshore advection in BBL reached the shoreward limit of the front thus shut down the offshore movement. During neap tide, the Ekman transport in the BBL of the plume front was landward (Figure 14c), which tended to push the plume front shoreward. It should be noted that this stable status may have not been reached during the neap tide since the BBL flow is onshore everywhere, even shoreside of the front. This is understandable because the neap tide period is too short for the plume to adjust to a new equilibrium state. However, when looking into along-shelf velocity profiles during spring tide (Figures 1c and 4d), one can see that half of the BBL of the plume front was occupied by down-shelf flow. This was due to the thick BBL that mixed down the down-shelf momentum from the upper layers into the bottom. In addition, increased tidal rectification in spring tide also enhanced the down-shelf momentum in the BBL. Such a down-shelf bottom flow produced offshore Ekman transport in the BBL (Figure 14c) and then moved the plume front seaward (Figure 14d). While during neap tide, because of the shrinking BBL thickness and the thermal wind balance, the trapping depth decreased. Under the upwelling-favorable summer monsoon, buoyancy in the ZMCW is depleted by the offshore and upstream wind-driven transports. Without a buoyancy supply, the plume front should disappear. This raises the question, what mechanism continuously provides the plume water for the ZMCW under upwelling-favorable summer monsoon conditions? It is plausible that the down-shelf current driven by the tidal-induced plume front can act as the buoyancy source of the ZMCW. To confirm the hypothesis, we tested the sensitivity of the buoyancy supply to the ZMCW on tidal forcing. Two numerical experiments (Exp7 and Exp8) were configured. Both of the two experiments ran from the same initial conditions and under the same boundary conditions as Exp3 for the first 120 days. After day 120, tidal forcing in Exp7 and Changjiang River discharge in Exp8 were turned off, respectively. The surface 31-psu isohalines, bottom salinity gradient, and residual surface currents on days 120, 160, and 190 in Exp7 and Exp8 are shown in Figure 16. Once tidal forcing was switched off (Exp7), the bottom-trapped plume in the ZMCW rapidly transformed into a surface-advected plume and retreated to the north. The down-shelf current in the ZMCW also disappeared. The transition of a bottom-trapped plume to a surface-advected plume under upwelling-favorable winds was in agreement with theoretical studies (e.g., Lentz, 2004; Pimenta & Kirwan, 2014; Whitney & Garvine, 2005). However, when tidal forcing was on (Exp8), despite upwelling-favorable wind forcing and the absence of runoff, the trapping depth of the plume in the ZMCW remained stable and the plume water did not immediately turn north. Also, there were still down-shelf currents presenting above the bottom front. The down-shelf volume transport and freshwater transport of the ZMCC (Figure 17) experienced a greater descending rate in Exp7 than in Exp8. Therefore, although river discharge ultimately controls the buoyancy supply to the whole plume, tidal forcing plays a more significant role in directing the buoyant water to the ZMCC.

For a tidal-induced plume front, river discharge is more like a buoyancy source than a momentum provider. To investigate how tidal forcing directs buoyant water to the ZMCC, another idealized experiment (Exp9), without river discharge but commencing with a stratified ocean, was tested. At the start of the simulation, the initial salinity field was set as follows: the salinity of the top 5 m was set to 0, salinity of water with
**Figure 14.** Sea surface elevation (a), BBL thickness (b), cross-shelf volume transport in bottom boundary layer (c), and tidal-averaged bottom salinity gradient (d), during spring and neap tides in Section A from Exp5 (mentioned in Figure 10). Solid lines indicate spring tide; dashed lines indicate neap tide. The periods of tidal averaging during spring and neap tides are shadowed within solid lines and dashed lines in (a), respectively. The x axis is set as water depth to better describe the trapping depth.

**Figure 15.** Tidal mean vertical eddy viscosity ($K_m$) and salinity profiles of Section A during spring and neap tides in February and August from Exp0. (a) Neap tide in February, (b) neap tide in August, (c) spring tide in February, and (d) spring tide in August. The logarithmic scale was used to plot $K_m$ profiles for better visualization. Red contours indicate salinity profiles.
depth greater than 10 m was set to 35 psu, and the salinity of water between 10- and 5-m depths was interpolated linearly. When simulating, all external forcings were excluded, with the exception of tidal forcing. The evolution of bottom salinity gradient, surface residual currents, and down-shelf volume transport in Section A are shown in Figure 18. After 60 days of the simulation, the tidal-induced front in the ZMCW stabilized where the Changjiang River plume front remained (Figure 18b). On day 120, because of the tidal dispersion and the lack of buoyancy supply, the tidal-induced front gradually dispersed (Figure 18c). The tidal-induced front also drove down-shelf currents above the bottom front and the transport was in the same order as the ZMCC (see residual surface currents in Figures 18a–18c and down-shelf transport in Figure 18d). Although Exp9 may exaggerate the amount of potential energy that the Changjiang River could supply, it suggested a self-sustaining process of the tidal-induced plume front in ZMCW. As long as buoyancy is continuously supplied by the Changjiang River, the current driven by the tidal-induced front will transport the buoyancy down-shelf to the ZMCW and maintain the plume front.

Moreover, it was found that most down-shelf transport in late spring and summer occurred during spring tides (Figure 11i). This contradicts nonwind numerical model results described in M. Li and Rong (2012), which suggest that down-shelf transport decreases when tidal mixing augments. As previously discussed, the summer ZMCC is sustained by the tidal-induced bottom-trapped plume front. SH values (Figure 13) show that the predicted locus of the tidal-induced front in the ZMCW has a spring-neap cycle, which is due to large fortnightly variations of tidal forcing. During neap tides, the trapping depth decreases due to the shrinking tidal mixed zone, and the plume front generally tilts and tends to transform to a surface-advected plume. As a result, the transport of the ZMCC decreases. During spring tides, the tidal mixed zone expands and the trapping depth increases, raising the transport of the ZMCC. Additionally, tidal rectification in spring tide induces an intensified down-shelf current on the top of the BBL, which further enhances the ZMCC (Figure 7). The largest difference between the results of the current study and those reported by M. Li and Rong (2012) relates to wind forcing. In M. Li and Rong (2012), wind forcing is excluded, and the plume remains bottom trapped. In their situation, the enhanced tidal mixing in spring tides suppresses the down-shelf advection...
of the plume. However, under upwelling-favorable winds, the plume tends to transform to surface-advected status and retreats in an up-shelf direction. Strong tidal mixing will sustain a bottom-trapped plume, which will, in turn, maintain down-shelf transport.

Therefore, in late spring and summer, the upwelling-favorable summer monsoon tends to push coastal plume water in the ZMCW offshore, which weakens the plume and creates stratification in this area. However, in shallow water, the water column remains homogeneous due to strong tidal mixing. As a result, a tidal front is constructed in the transition area between well-mixed coastal water and stratified offshore water. Due to the baroclinic pressure gradient, the tidal-induced front stimulates a down-shelf current, that is, the summer ZMCC. The tidal rectification in spring tide, induced by strong tidal forcing on the sloping bathymetry and enhanced by stratification, further augments the down-shelf current. Freshwater from the upstream plume is transported to the ZMCW via the down-shelf ZMCC, which supplies buoyancy to the plume in the ZMCW and then maintains the stratification. The tidal-induced front in stratified water then enables the ZMCC to persist under the upwelling-favorable summer monsoon.

As many river plumes experience intense tidal mixing, this study provides an alternative explanation on the formation of bottom-trapped plumes and associated buoyant coastal currents on energetic continental shelves. For example, Guo and Valle-Levinson (2007) also found that the tidal mixing essentially changes the vertical structure of the Chesapeake Bay outflow plume and thus restricts the up-shelf extension and augments its down-shelf transport. The Rhine ROFI’s structure, stratification, and frontal propagation speed are under modulation of tidal forcing, through tidal straining, advection, and mixing, exhibiting semidiurnal and fortnightly cycles (Simpson et al., 1993; Simpson & Souza, 1995; Fisher et al., 2002; De Boer et al., 2008; Rijnsburger et al., 2018, etc.). The Changjiang River plume, Chesapeake Bay outflow plume (Guo & Valle-Levinson, 2007), and the Rhine ROFI are generally over relatively shallow shelves, and therefore, the interaction between tidal forcing and bottom topography is strong. The mechanism proposed in this study thus may work. However, for cases like Columbia River plume, because of the steep shelf, the interaction
between bottom topography and river plume is weak. Hence, although the plume is strongly affected by the tide near the river mouth (MacCready et al., 2009; McCabe et al., 2009; Nash & Moum, 2005), it is governed by other processes far beyond this tidal region (Hickey et al., 2010; Horner-Devine et al., 2009). This means that the YC97 theory is still instructive. If a river plume was predicted by YC97 to be a surface-advected one, it likely remains so under the tidal effect, due to the loose contact between plume water and bottom topography. However, if a river plume was predicted to be bottom-advected one, the strong tidal mixing over sloping topography may augment this tendency.

5. Conclusions

In this study, the characteristics of the Changjiang River plume in the ZMCW were investigated using a numerical model. The model was validated by data from eight cruise surveys conducted in different seasons. The model performed well in both surface plumes and bottom plumes. The model results revealed that although the expansion of the surface plume varies dramatically between seasons, the bottom plume in the ZMCW persists in the coastal area. The bottom front persists between the 20- and 40-m isobaths with a width of ~20 km all year-round, which indicates that the plume is a bottom-trapped plume. A strong surface-enhanced down-shelf geostrophic coastal current is induced by the baroclinic pressure gradient of the plume front, which is known as the ZMCC. The maximum down-shelf velocity exceeds 0.4 m/s, and the center of the coastal current is located over the bottom front.

The ZMCC was conventionally thought to be a wind-driven coastal current, prevailing in winter and disappearing in summer. However, P. Li et al. (2014) reported that a down-shelf current can also be observed in the ZMCW in summer, and Wu et al. (2013) suggested that buoyancy plays an essential role in the water transport of the ZMCC. Model results obtained during the current study are in agreement with those findings. The wind-driven component accounts for less than 23% of the annual down-shelf volume transport of the

Figure 18. Bottom salinity gradient and residual surface currents in days 2, 60, 120 (a–c) and residual down-shelf volume transport across Section A (d) from Exp9.
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