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DYNAMICS OF RHODE ISLAND COASTAL WATERS

BY

QIANQIAN LIU

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE

REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

IN

PHYSICAL OCEANOGRAPHY

UNIVERSITY OF RHODE ISLAND

2015

DOCTOR OF PHILOSOPHY DISSERTATION
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ABSTRACT

The overall objective of this dissertation is to establish a framework for the dynamical understanding of a number of circulation features found in Rhode Island coastal waters. Observations reveal a cyclonic, near-surface circulation around the periphery of Rhode Island Sound (RIS) that occurs in summer stratified conditions and disappears in winter when solar insolation and wind stirring result in strong vertical mixing. The dynamics of this circulation is investigated through a series of numerical simulations. We attribute the summer intensification of this ‘periphery current’ to a continuous topographically rectified tidal residual current, and to a circulation produced by seasonal deep thermohaline gradients often observed from May to September.

The depth-averaged vorticity is found to be most useful for diagnosing the sources of the topographically-rectified tidal residual current, and that diagnostic is also applied toward our understanding of another feature of Rhode Island coastal circulation: a pair of opposite sign headland eddies that are found around Montauk Point in the Block Island Sound (BIS). The RIS cyclonic residual current is attributed to the nonlinear transfer of vorticity generated by the Coriolis torque when tidal currents oscillate, which is opposed by bottom frictional torque associated with topographic gradients and velocity shear. In addition to bottom frictional torque, we show that the nonlinear advection of vorticity is important in generating the eddy pair in BIS. The effect of bottom thermal fronts on the circulation is illustrated through process-oriented, reduced-physics experiments by varying the surface solar radiation.

Other important features of the Rhode Island coastal circulation are found in the region between BIS and Long Island Sound (LIS), including a buoyancy-driven coastal current along the southern shore of LIS and periodically detached freshwater patches to the south of Montauk Point and in central BIS. Buoyant discharge coming out of LIS, mainly from the Connecticut River, forms a seasonal plume outside BIS between the coast of Long Island and the denser shelf waters. The plume's seasonal variability and its response to tides, winds and surface heating are investigated through a series of process-oriented numerical experiments. In winter and spring, the plume is intermediate with a large surface offshore extension detached from the bottom, while in summer, the front is bottom-advected with most of the width in contact with the bottom, featured by steep isopycnals. Strong summer insolation together with weak buoyant discharge and weak winds generates the narrowest and steepest summer plume. In addition, small changes in tidal currents over the spring-neap cycle cause significant, monthly fluctuations in turbulent mixing and vertical stratification in central BIS, modulating the freshwater incursions that generate episodic freshwater patches moving downstream along the southern shore of Long Island and towards RIS in the gap between Block Island and Point Judith. Observational evidence of these detached patches of fresh water is also discussed in this study.

ACKNOWLEDGMENTS

I would like to express my sincere gratitude to my advisor, Dr. Lewis M. Rothstein, for his support and patient guidance. Without his encouragement and support, I cannot finish my Ph.D. study and this dissertation would not have been possible. I acknowledge Dr. Yiyong Luo for his persistent help in my research. I acknowledge Dr. David Ullman and Dr. Dan Codiga for providing my research with observational evidence and for the useful and insightful discussions, and Dr. David Ullman and Dr. Christopher Kincaid and Dr. Jason Dahl for serving in my committee and giving me useful instructions. I acknowledge my husband, Bin, for his unconditional love and accompany. I acknowledge my parents and my brothers for their encouragement to help me undergo the hard times. Finally, I dedicate this dissertation to the memories of my beloved grandmother and grandfather.

PREFACE

This dissertation is in manuscript format. It is composed of two manuscripts. The first one will be submitted to the Journal of Geophysical Research and the second will be submitted to the Journal of Physical Oceanography for publication.

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Manuscript 1

Formation of a Cyclonic Current Along the Periphery of Rhode Island Sound

by

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In preparation for submission to Journal of Geophysical Research

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Abstract

The dynamics of the seasonal cyclonic circulation around the periphery of Rhode Island Sound (RIS) is investigated through a series of numerical simulations and theoretical constructs that are then compared against all available observations. We attribute the summer intensification of the periphery current to a combination of year-round topographic rectification of the tidal residual current and weak summertime winds, that then establishes a thermal wind balance with the seasonally varying deep thermohaline gradients often observed from May to September. A depth-averaged vorticity balance is used to diagnose the sources of the topographically-rectified tidal residual current in RIS, and these diagnostics are also applied to a pair of opposite sign headland eddies around Montauk Point in Block Island Sound (BIS) that is the other major circulation feature found in the model. The RIS cyclonic residual current is due to the nonlinear transfer of vorticity generated by the Coriolis torque when tidal currents oscillate, which is opposed by the bottom frictional torque associated with the topographic gradient and velocity shear. In addition to previously emphasized role of bottom frictional torque and ‘wall’ boundary friction, we find that planetary vortex-tube stretching is also important, especially for the BIS headland eddies. The balance provided to the circulation by deep thermal fronts are illustrated with two process-oriented experiments with strong and weak short radiation, and by a geostrophic calculation. We believe that these dynamics are operating elsewhere in the world’s coastal oceans, and propose generalized criteria for determining where one might find these circulations.

1. Introduction

Rhode Island Sound (RIS), a semicircular embayment off the southern Rhode Island and Massachusetts coast, is located offshore of the mouth of Narragansett Bay and bounded to the west by Point Judith and Block Island (BI) and to the east by Martha's Vineyard and the Elizabeth Islands (Figure 1-1). Waters in RIS are connected with Block Island Sound (BIS) through the gap between Point Judith and BI. Based on field measurement, Kincaid et al. (2003) identified a cyclonic, near-surface circulation around the periphery of RIS that is usually present in summer, stratified conditions that then disappears in winter when weak solar insolation and wind stirring results in strong vertical mixing. In summer, the cyclonic circulation moves along the coast of Rhode Island, flowing north into RIS from the southeast, to the west at the inner shelf where it encounters potentially significant dynamical exchange processes associated with Narragansett Bay, and then to the southwest in the western portion of RIS (see Figure 1-2 for a schematic of the circulation). A seasonal thermocline accompanies this circulation in spring and summer (Shonting and Cook, 1970). During the winter months this entire flow pattern collapses, with much weaker currents and different trajectories (Ullman and Codiga, 2010).

Luo et al (2013) find that tidal residual effect is a significant dynamical contributor to the circulation in RIS but they did not detail the dynamics that give rise to the strong tide-generated cyclonic gyre around RIS. Pingree (1978) and Zimmerman (1981) introduced a novel approach to understanding the mechanisms of tidal residual circulations by examining the vorticity of tidal flows, and Robinson (1981) provided further insights based on the transfer of tidal vorticity to the mean

field. Moreover, Robinson (1981) illustrated the unique role of quadratic bottom friction when there was a topographic slope perpendicular to the tidal residual currents. To fully understand the sources of tidal residual circulations, Ridderinkhof (1989) used the depth-averaged vorticity equation to show that the essential balance was provided by the advection and dissipation of tidal vorticity generated by column stretching and squeezing, and by the frictional torque due to depth gradients and velocity shear.

In addition to the tidal residual current, another competing hypothesis for the cyclonic circulation in RIS finds a thermohaline circulation that is in thermal wind balance with sharp horizontal fronts of bottom temperature and salinity that are usually present in spring and summer. The horizontal fronts are produced by the spatially varying distribution of tidal mixing (Hill et al., 2008), which has been found in many other shallow tidal areas in the Northern Hemisphere (Horsburgh et al., 2000).

In addition to the periphery current, a pair of eddies around Montauk Point was present in the numerical simulations of both Edwards (2004) and Luo et al. (2013). Similar closely-connected headland eddy pairs have been found in other regions and their dynamics have been attributed to many different mechanisms, with bottom frictional torque (Pingree, 1978; Geyer and Signell, 1990) and no-slip lateral ‘wall’ boundary friction proposed by Zimmerman (1981) being the most popular. In Zimmerman's (1981) mechanism, the velocity gradient normal to the wall is associated with the wall's lateral frictional boundary layer that is then transported to the mean field from the oscillating tidal field by the nonlinear advection of vorticity. Based on

the analysis of the transport vorticity equation around an idealized, symmetric promontory in a numerical simulation, Park and Wang (2000) attributes the eddies to the transfer of topographic vorticity, which is consistent with Robinson's (1981) interpretation. In essence, the side-wall boundary works the same way in generating the velocity gradients as a slope that gradually shoals around a promontory (Zimmerman, 1981). In BIS, the topographic features make the situation rather more complicated. Edwards (2004) has emphasized the importance of vortex stretching and tilting for the anticyclonic eddy south of Montauk Point without, however, connecting it with the cyclonic eddy to the north of Montauk Point.

The purpose of this study is to examine the sources of the topographically rectified residual currents in RIS and BIS, and to investigate the role of both tidal residual currents and deep thermohaline gradients in the summer circulation in RIS through both a series of three-dimensional numerical simulations and theoretical analysis. The numerical model configuration, design and model validation are described in Sections 2 and 3. The tidal rectification in RIS and BIS are discussed in Section 4, and the contributions from bottom thermal gradients are presented in Section 5. A summary and a discussion for generalizing these results beyond Rhode Island's coastal waters are given in Section 6.

2. Model Configuration

The numerical simulations are performed using the Regional Ocean Modeling System (ROMS; Shchepetkin and McWilliams 1998, 2003, 2005; Haidvogel et al. 2000) that employs a terrain-following coordinate for the vertical. The high resolution ‘local’ domain covers RIS, BIS, Long Island Sound (LIS) and the adjacent inner shelf area, with a uniform horizontal grid of 800 m and 30 terrain-following vertical layers, and this domain is one-way nested within a ‘regional’ domain with a constant horizontal resolution of 5 km (Figure 1-3). Compared with the former configuration used in Luo et al. (2013), we have improved the resolutions over Narragansett Bay to better capture its dynamical exchanges with RIS, and increased the number of vertical layers from 15 to 30 to better resolve the vertical structure, especially in the bottom boundary layer.

Our ‘realistic’ experiment (REAL) is driven by atmospheric forcing, observed river discharges and tides in an effort to simulate the seasonal variability of RIS circulation. Daily averaged climatological surface atmospheric fluxes (including winds, air temperature, air pressure, relative humidity, rainfall rate, short wave, and long wave radiations) from 2004 to 2012 were obtained from the North American Regional Reanalysis (NARR; Mesinger et al. 2006). Bulk formulas are applied to calculate latent and sensible heat fluxes.

Daily river discharge data for the major rivers (from north to south - Taunton River, Blackstone River, Pawtuxet River, Pawcatuck River, Connecticut River, and Housatonic River) were obtained from the United States Geological Survey (USGS;<http://waterdata.usgs.gov/nwis>), and we re-scaled these to account for the

ungauged portions of the watershed to improve the salinity in LIS and BIS (Zhang et al., 2010). This re-scaling has improved the discrepancy between model salinity results and observations from a maximum difference of ~ 3 PSU (using un-scaled data) to within 0.5 PSU. REAL utilizes 5 major tides (M_2 , N_2 , S_2 , O_1 and K_1), which represent 80% of the variance of the total depth-averaged tidal kinetic energy as obtained from the Advanced Circulation Model for Oceanic, Coastal and Estuarine Waters (ADCIRC) tidal simulation (Luettich et al., 1992).

To study the topographically-rectified tidal current, we configured a hierarchy of process-oriented experiments, specifically targeted at understanding the fundamental dynamics of tidal rectification mechanisms associated with topography, Coriolis force, and momentum advection. To simplify the problem, only the M_2 tide, which accounts for 85% to 90% of the energy from the five major constituents, was retained in all of these experiments. Because of the unique role played by quadratic bottom friction, the quadratic bottom drag formulation was used in all tidal simulations with a drag coefficient of $C_D = 7.5 \times 10^{-3}$. These process-oriented experiments are defined as follows: ‘RealM2’ - realistic bathymetry, with both Coriolis force and momentum advection retained; ‘Flat’ – ‘RealM2’ but with a flattened bottom; ‘NoRotate’ – ‘RealM2’ but non-rotating; and ‘NoAdv’ – RealM2 without momentum advection.

Two additional process-oriented experiments are set to explore the contribution of bottom intensified thermohaline gradients for balancing the summer-intensified circulation. The first (‘HeatS’) is driven by climatological mean surface heating in August and starts from a homogenous rest state. The second (‘HeatW’) is driven with

half the solar radiation of ‘HeatS’. We run both experiments for three months and examine the time-averaged results in the third month.

3. Model Validation

Before performing an analysis of the model results, we validated our simulations through comparisons with all available observations. The model tides are compared to observed tidal ellipses (Mau et al., 2007; Moody et al., 1984) and Coastal Ocean Dynamics Applications Radar (CODAR; Ullman and Codiga, 2004; Luo et. al., 2013). We also validate model tidal sea level variations with observations from bottom-mounted acoustic Doppler current profilers (ADCP) at stations PO-S and PO-F, whose locations are marked in Figure 1-1 (Ullman and Codiga, 2010; Figure 1-4). Root mean square errors for sites PO-S and PO-F are 0.16 m and 0.04 m, respectively.

We next compare the monthly mean, depth-averaged model currents with moored observations from 2009 and 2010 (Figure 1-5) that are a combination of data from the Rhode Island Ocean Special Area Management Plan project and the project funded by RISG and RIEP (Christina Wertman in URI is working on the analysis of the data). In May, both the model and the observations show relatively weak (~ 2 cm/s) onshore velocities in RIS, with stronger currents (~ 8 cm/s) located to the south of Block Island in a region that connects RIS and BIS. In July, the model successfully captures the westward jet offshore of Narragansett Bay as well as the southwestward current leaving RIS that appears in the observations. With this, the model does a very good job in the center and south of RIS in spring and summer.

On the inner shelf of RIS, close to the mouth of Narragansett Bay, the model captures the seasonal variability of circulation and water exchange between Narragansett Bay and RIS that has been observed by acoustic Doppler current profiler surveys between April 1998 and July 1999 (Kincaid, 2003; Table 1). Consistent with the surveys, during summer, a relatively strong transport of $\sim 4000 \text{ m}^3/\text{s}$ enters the survey area through the Brenton Reef line; weaker transports flow into the survey region through the East Passage and West Passage lines (please see Figure 1-1 for the location of these lines). However, most of the transport into the region comes through the Narragansett Beach line, with a relatively smaller transport through the South Line. During winter, water exchange over the survey region changes significantly with a much-diminished transport through the Brenton Reef, Narragansett Beach and South Line, even reversing direction through the Brenton Reef and South Line in January and February. Compared with the observations, in winter, the transport through the West Passage is underestimated; in summer, the model produces a much smaller transport through Narragansett Beach and a larger transport through the South Line. This discrepancy may be due to the model's treatment of rivers in Narragansett Bay, in which we include all river input to Narragansett Bay from the three major rivers but ignore underground water discharges that have proven to be important to Narragansett Bay (Pilson, 1985). Nevertheless, the correspondence in seasonal variation of transport and monthly mean currents is quite encouraging and justifies using the model to study the dynamics of the circulation in RIS.

4. Tidal Rectification

4.1 Process-oriented Experiments

A major contributor to the periphery current in RIS is the tidal residual circulation (Luo et al, 2013). In this study we configure a hierarchy of process-oriented experiments to understand the role of topography in setting this circulation. Figure 1-6 shows the depth-averaged tidal-rectified residual currents for experiments a) RealM2, b) Flat, c) NoRotate, and d) NoAdv. In each case the currents are averaged for the last month of a 3-month simulation. For all experiments we select a quadratic bottom friction formulation with a bottom drag coefficient of 7.5×10^{-3} . For the RealM2 experiment the tidal rectification in RIS is characterized by a cyclonic circulation of ~ 3 cm/s, originating from the south of Martha's Vineyard, moving southwestward toward Block Island and eventually joining the circulation around Block Island. In BIS and LIS, the residual current forms an eddy pair around Montauk Point, between which there is an eastward current that is larger than 5 cm/s. This eddy pair pattern has also been found in different regions (e.g., Pingree, 1978; Geyer and Signell, 1990). In addition, a weak cyclonic eddy is present to the east of the strong anti-cyclonic eddy (southwest of Block Island). Because of the complex regional geography and topography and stronger tidal currents, the generation mechanism(s) that give rise to these eddies is rather complex and will be discussed in detail later in this manuscript.

Figure 1-7 shows the relative vorticity in the region of interest. Residual eddies in RIS includes positive relative vorticity eddies inside most of RIS, negative relative vorticity eddies along the coast where currents encounter irregular promontories, and

negative relative vorticity eddies in the west of RIS. The positive relative vorticity eddies inside RIS are set by a residual current that moves cyclonically inside RIS and along its periphery. Flow moving out of RIS to the east of BI identified with the negative relative vorticity eddies in the west of RIS. In the southeast of RIS, around Gay Head, a pair of oppositely signed eddies occurs, with the positive cyclonic eddy to the west and anticyclonic eddy to the east (inside the green square in Figure 1-7). As an important feature generated by tides, this pattern has been captured by both observations (Geyer and Signell, 1990) and numerical simulations (Park and Wang, 2000), and has been ascribed to a balance between vorticity advection and topographic vorticity.

The second experiment, Flat, is a flat bottom simplification of experiment RealM2. Comparing the circulations between RealM2 and Flat (Figure 1-6) shows rather conclusively that the major tidal residual current patterns in RIS and BIS are mainly topographically-rectified. For example, due to the absence of bottom topography the eddy pair around Montauk Point is significantly weaker, as is the cyclonic eddy south of the eddy pair. Moreover, the anticyclonic eddy around BI and cyclonic circulation around RIS almost disappears, leaving a relatively sluggish westward flow in the northeast region of RIS. We therefore conclude that topography is the dominant contributor to the residual current in the study region and will focus on the dynamics of topographic rectification in the following discussions.

The experiment NoRotate emphasizes that we cannot ignore the role of the Coriolis force for dynamical understanding of the periphery current in RIS. In contrast to RealM2, NoRotate does not produce an anticyclonic circulation in RIS; instead it

generates (weak) negative relative vorticity over most of RIS (Figure 1-8). The current flowing out of Vineyard Sound, the source water for the periphery current, flows directly offshore out of the sound instead of diverting to the right of the current direction as would be the case with rotation. We therefore conclude that, in addition to topography the Coriolis force needs to be included in our dynamical understanding of the periphery current, consistent with the studies of Huthnance (1973) and Loder (1980).

Even though an eddy pair is generated in the west of BIS in experiment NoRotate, as compared with the eddy structure generated by Realm2 the entire paired eddy structure is tilted cyclonically and is weaker. In addition, less water is trapped in the eddy north of Montauk Point, with stronger flow passing through the gap between Point Judith and Block Island, contributing to a broader anti-cyclonic circulation east of Block Island. In addition, the weak cyclonic eddy southwest of Block Island is smaller and weaker. Considering the differences in these eddy structures between experiments Realm2 and NoRotate, we can conclude that mechanisms associated with the Coriolis force is also important for dynamical understanding of the eddies in BIS. On the other hand, the fact that completely omitting the Coriolis force does not eliminate the eddies proves that other mechanisms independent of earth rotation but associated with bottom slope apparently play more important role, which motivates a detailed analysis of the vorticity balance in the next section.

To conclude this set of experiments, we turn off momentum advection (NoAdvect) and obtain a very simple and weak circulation without any of the major circulation patterns discussed above. According to Pingree (1978) and Ridderinkhof

(1989), the essential tidal rectification process is the advection and dissipation of vortices generated by periodic tidal currents, i.e. it is advection that transfers vorticity from the periodic tidal currents to the residual circulation. Hence no obvious residual current is generated without advection terms.

4.2 Process-oriented Experiments

We follow Robinson (1981) and Ridderinkhof (1989) by using the flow vorticity as the dynamical foundation of our analyses. The vorticity budget, after averaging over a tidal cycle, can be written as:

$$\begin{aligned} \underbrace{\overline{\frac{\partial \omega}{\partial t}}}_a = & - \underbrace{\left(u \frac{\partial \omega}{\partial x} + v \frac{\partial \omega}{\partial y} \right)}_b - \underbrace{\overline{\omega \nabla \vec{u}}}_c - \underbrace{\overline{f \nabla \vec{u}}}_d - \underbrace{\overline{\frac{k \sqrt{u^2 + v^2}}{H^2} \vec{u} \times \nabla h}}_e \\ & + \underbrace{\overline{\frac{k}{H} \vec{u} \times \nabla (\sqrt{u^2 + v^2})}}_f - \underbrace{\overline{\frac{k \sqrt{u^2 + v^2}}{H} \omega}}_g + \underbrace{\overline{v \nabla^2 \omega}}_h \end{aligned}$$

where the overbar denotes a time average (over one tidal cycle), $\vec{u}$ (u, v) are the vertically integrated velocity components; H is the sum of the local water depth (h) and the local tidal elevation above mean sea level; ω is the relative vorticity of the depth-averaged velocity; f is the Coriolis parameter; k is the (constant) bottom friction coefficient; and v is the horizontal eddy viscosity. The vorticity budget can be addressed with respect to the following individual terms: the local (a) and advective (b) changes of relative vorticity; stretching of the water column by relative (c) or planetary (d) vorticity; frictional torques due to the sloping bottom obliquely transverse to the tidal currents (e) and the horizontal velocity shear (f); and the damping of vorticity by bottom friction (g) and horizontal diffusion (h). Term a is

identically zero after averaging over a tidal cycle. Terms e, f and g all originate from the bottom friction. The term f only occurs when a quadratic bottom friction formulation is used. The term h is found to be small in all of our experiments.

We begin our dynamical diagnoses of the currents in RIS and BIS by calculating each of these terms. Figure 1-9 shows the terms advecting (b), producing (c, d, e, f) and dissipating (g) vorticity averaged over 20 tidal cycles of hourly numerical data. Consistent with the process-oriented experiments and available studies (for example, Pingree, 1978), advection of vorticity, which is responsible for the nonlinear transfer of vorticity from periodic tidal currents to the residual circulation, is essential for both RIS and BIS. Advection is negative (positive) to the south (north) of Montauk Point and its effect is comparatively weaker in RIS than in BIS, but still dominates the residual circulation along the east coast of RIS, especially the circulation around Gay Head as described by Park and Wang (2000). In contrast, relative vorticity stretching is negligible, especially in RIS, while in BIS it can generate weaker vorticities with negative values over shallow regions (to east of Montauk Point) and positive values over the deep waters (to north and south of Montauk Point).

Planetary vortex stretching creates positive eddies along the inner shelf of RIS, while frictional torques associated with bottom slope and velocity shear counterbalance the effects of the Coriolis torque. In BIS, the complicated geography makes the contributions from different source terms more complex than the available studies that are configured with idealized, symmetric promontories (Park and Wang, 2000). In addition to the essential contribution of vorticity advection, Figure 1-9

shows that planetary vorticity stretching creates positive eddies to the east of Montauk Point. This result is consistent with Edwards' (2004) study, in which vortex stretching is found non-negligible within a region 5 km offshore of Montauk Point. However, our work differs from Park and Wang (2000) who suggest that the frictional torque associated with bottom slopes were canceled out by the frictional torque associated with velocity shear. Our simulation around Montauk Point shows these two kinds of frictional torques reinforce each other; in BIS, they both generate positive eddies over the shallow waters to the east of Montauk Point and negative eddies over the deep waters to the north and south of Montauk Point. Generally, the pair of opposite gyres is generated by a combination of vorticity advection, relative vorticity stretching, frictional torque, and other non-topographically related mechanisms, with the largest contribution from nonlinear advection.

Taking guidance from both Robinson (1981) and Zimmerman (1981), we construct a simplified, interpretive schematic to analytically demonstrate: a) the production of vorticity by the Coriolis torque and the nonlinear transfer of vorticity, i.e., terms b and d , (Figure 1-11) and; b) production of vorticity by topographic slope torque and nonlinear transfer, i.e., terms b and e , (Figure 1-12). To analyze the transfer of vorticity from periodic tidal currents to the residual circulation during different tidal stages, hourly snapshots of depth-averaged, periodic currents are shown in Figure 1-10. We refer to the stage when tidal currents move from the outer shelf to inner shelf, i.e. from the 1st to the 5th hour, as flood, and the stage with opposite direction, i.e. from the 7th to 11th hours, as ebb.

Tidal currents in BIS are typically strong and parallel to isobaths most of the time (figure 1-11a) but there are times when these currents obliquely move across slopes (Figure 12a; south of Montauk Point around the 1st and 7th hours). The oblique across slope direction results in a non-zero cross product between the bottom slope and the tidal current and is important for vorticity generation associated with frictional torque, that is, term e (further discussed below). In RIS, tidal currents have a component that is perpendicular to the topography most of the time. Furthermore, except for the 1st and 7th hours, tidal currents obliquely cross the slope in the northeast of RIS (close to the mouth of Buzzards Bay) west to east during flood and east to west during ebb.

According to Zimmerman (1981), the nonlinear transfer of vorticity from oscillating tidal currents to a residual circulation is represented by a “tidal stress”, which is the integral of the tidal mean flux of vorticity along a streamline, $\int \overline{\vec{\omega} \times \vec{u}} dl$, where l is a residual streamline, $\vec{u}$ is the instantaneous tidal velocity at a point of l , and $\vec{\omega}$ is the vorticity of the tidal current velocity field.

Both the process-oriented tidal experiments and our analyses of the depth-averaged vorticity equation suggest that the Coriolis torque is important in RIS. In BIS the idealized simulation and the vorticity balance indicate that the influence of the Coriolis torque is small compared with the other terms. Note in the following discussions, the generation of vorticity is considered in the environment of “tidal stress” instead of the time- and depth-averaged vorticity balance shown in Figure 1-9, which is only used to help us assess the relative importance of each term. The reason is that in the depth-averaged equation, all the production terms are isolated from the

nonlinear advection that is essential to tidal rectification. When we don't consider nonlinear advection, if one term has opposite values during flood and ebb, averaged over a tidal cycle, the two stages annihilate each other. Only when combined with the nonlinear advection, the terms become physically meaningful.

We first focus on the situation in RIS (Figure 1-11b) where a highly simplified bathymetry is represented by two close, dotted lines. During flood tide, when a water column moves from deep water to shallow water it experiences stronger Coriolis force on the shallow side, and thus an anti-cyclonic Coriolis torque. The onshore tidal current brings the anti-cyclonic vorticity out of the streamline. During ebb tide, when the water column moves in the opposite direction, there is a cyclonic Coriolis torque exerted on the water column and therefore a cyclonic vorticity brought into the streamline by the offshore current. Because both flood and ebb processes involve same sign of “tidal stress”, when averaged over a tidal cycle, the tidal currents generate a net influx of cyclonic vorticity in the deep region and an outflux in the shallow region, giving rise to a cyclonic circulation along the isobath.

When tidal currents obliquely cross topographic slopes, a tidal residual circulation is generated by frictional torque, and the circulation direction is determined by the angle between the oscillating tidal currents and the slopes. In RIS, we simplify the flood and ebb stages (Figure 1-12c). Because of mass conservation, a water column moving across the slope will experience stronger bottom friction on its shallower side, so that, during flood when the column moves in the directions as shown in Figure 1-12c, a cyclonic frictional torque is generated. The flood stage transports cyclonic vorticity out of the deep water and into the shallow water. During

ebb tide, an anticyclonic torque is produced by the bottom friction, and therefore an anticyclonic vorticity is brought out of the shallow water and into the deep water. Averaged over a tidal cycle, the frictional torque tends to produce an anticyclonic circulation along the isobath to compete with the cyclonic circulation produced by the Coriolis torque.

The generation mechanism around Montauk Point is similar with RIS but more complicated due to its complex topography and tidal currents. A highly simplified schematic is shown in Figure 1-12b. Because currents cross bottom slopes in different directions before and after passing Montauk Point, the shallow water on the east of Montauk Point is divided into two regions. During flood tide, when a water column encounters the isobaths between the south shallow water east of Montauk Point and the deep water to its south, it experiences stronger bottom friction on the shallower side and therefore a cyclonic frictional torque. The flood stage transports cyclonic vorticity out of the deep water to the south of Montauk Point. Similarly, after passing Montauk Point, when the water column moves from the north shallow water to the deep water located north of Montauk Point in the direction shown in Figure 1-12b, a cyclonic vorticity is brought into the deep water. Over the shallow regions, the change of tidal current direction annihilates the transport of vorticity generated before and after the turning point.

During ebb tide, when the water column moves in the opposite direction (i.e. from the deeper water northeast of Montauk Point through the shallower region east of Montauk Point and then into the deeper water southeast of Montauk Point), there is an anticyclonic vorticity flux out of the deep region to the north of Montauk Point and an

anticyclonic vorticity flux into the deep water south of Montauk Point. Overall, in an M2 cycle, cyclonic eddies are generated over the deep water to the north of Montauk Point and anticyclonic eddies over the deep water to the south of Montauk Point.

The frictional torque associated with a velocity gradient, namely ‘speed torque’, is discussed in detail by Zimmerman (1981) using both the processes of residual basin circulation and headland eddies. Our modeling results are consistent with his discussion by generating negative eddies over RIS through term f . In this term, friction generates a velocity gradient by diminishing velocities toward the lateral wall. Over BIS, its effect reinforces that of the slope torque. The weak eddy-pair around Montauk Point in Flat experiment is due to the slope torque. Note that this term is due entirely to the quadratic formulation of bottom friction and its magnitude is smaller than slope torque. We can conclude that the application of either a quadratic or linear bottom drag formulation will not affect the production of the major residual patterns, which has been proved by an idealized experiment with linear bottom drag coefficient. Transient depth-averaged vorticity terms and corresponding tidal currents are needed to validate the interpretive generation mechanisms.

According to Robinson (1981), topographic features having the length scale of the tidal excursion, $L_t = 2U/\omega$, are likely to have the greatest influence on the generation of mean flows by the tides. The tidal excursion in our study region is about 5-8 km. This explains why around Montauk Point and on the inner shelf of RIS topographically-rectified currents are more important than on the offshore regions where topographic features have larger length scales.

5. Thermohaline Gradients

5.1 Seasonal Variability

In coastal seas, a transition between the stratified, offshore region and the tidally mixed, onshore region often generates bottom and surface thermal fronts (Hill et al., 2008; Takeoka et al., 1997). Simpson and Hunter (1974) used an empirical criterion, H/U^3 (where H is the water depth and U is the root-mean-square value of depth-averaged tidal velocity), to estimate the locations of tidal mixing fronts. The criterion describes the energy balance between the demand for potential energy for a fully mixed condition and the supply of vertical mixing energy from the tidal current, with large values representing efficient vertical mixing.

Figure 1-13 plots the $\log_{10}(H/U^3)$ for M_2 tidal currents in RIS and BIS. It shows intense mixing over BIS and eastern and western RIS. In BIS, because buoyancy is also due to the freshwater entrained out of LIS and not just the tidal flow, this diagnostic does not capture the full story of frontal structure. But in RIS, the freshwater contribution to the dynamics is trivial (Ullman and Codiga, 2010), and it is therefore good to use this criterion to predict the locations of thermal fronts.

Seasonally averaged surface circulations in summer and winter from our REAL experiment are shown in Figure 1-14a. Consistent with the limited observations (Codiga and Rear, 2004; Codiga and Ullman, 2010), the summer circulation features a cyclonic circulation within RIS originating from the southwest of Martha's Vineyard, moving along the coast of Rhode Island, and flowing out of RIS to the southeast of Block Island before joining the southwestward downshelf flow in BIS. The circulation

pattern changes dramatically in winter when the cyclonic pattern in RIS essentially collapses, especially in the middle and eastern regions of RIS. The current out of BIS steers offshore, however its magnitude does not change as significantly as RIS over the seasonal cycle.

The seasonal variations are also reflected at depth as shown in Figure 1-14b. The section is from Block Island to the east of RIS (black line in Figure 1-13), which represents a typical physical profile for RIS. During summer, when strong solar insolation occurs, the transitions between the tidally mixed regions (in eastern and western RIS) and the stratified region in the interior of RIS are shown by the robust deep thermal fronts at the ends of the section with weak thermal fronts near the surface. Accompanying the deep thermal fronts is a strong northward jet along the eastern portion of the section and a strong southward jet along the west, defining the cyclonic circulation around RIS. During this time period, salinity remains much less stratified at both inner shelf and outer shelf with a spatial change of $< 0.2 \text{ PSU}$ over this region,

During winter, with the significant decrease of insolation and an increase of surface torque due to an increase in the seasonally varying winds, both temperature and salinity are vertically homogenized over most of the region. At the same time, currents in the top layers turn offshore and diminish to $< 1 \text{ cm/s}$ except for the region within 10 Km off Block Island that was influenced by the flow from BIS through the gap between Block Island and Point Judith.

5.2 Bottom Thermal Fronts in Idealized Experiments

We examine the impacts of the bottom and surface thermal fronts on the summer circulation by analyzing the results of experiments HeatS and HeatW. Monthly mean surface circulations from these two cases are shown in Figure 1-15a. Consistent with the REAL simulation, currents in both experiments feature a cyclonic circulation around RIS. Differences in the surface currents (HeatS-HeatW; Figure 1-16a) give a circulation difference pattern in the cyclonic direction in RIS, demonstrating that a surface heating reduction slows the cyclonic circulation around the periphery of RIS by ~ 5 cm/s. In the interior of RIS, the circulation difference is negligible. Between the cyclonic circulation difference and the interior of RIS, there is a weak (~ 1 cm/s) anticyclonic difference (Figure 1-16a), corresponding to a wider cyclonic jet in experiment HeatW than in HeatS.

Figure 1-15b shows the cross-sectional structures of temperature (black contours) and velocity (colors) across the section defined in Figure 1-13. At the ends of the section where $\log_{10}(H/U^3) \leq 4.2$ (Figure 1-13), water columns are vertically mixed by tidal mixing. In the interior where $\log_{10}(H/U^3) \geq 5$, stratification is strong due to lack of mixing. Temperature contrasts between the vertically mixed water columns and the deeper, stratified water column form a prominent bottom thermal front and a less prominent surface front. By thermal wind balance, the bottom thermal front drives a cyclonic circulation at the surface, while the surface thermal front tends to generate an opposite circulation (Figure 1-17). Thus the surface cyclonic circulations in Figures 1-14a and 1-15a emphasize the importance of the bottom thermal front.

Given that different pools of water along the cross-shelf direction respond to surface heating reductions to different extents (Hill et al., 2008), changes in thermal fronts are expected as follows. Because surface heating penetrates deeper in inshore, shallower waters than in offshore, deeper waters, a temperature reduction for inshore waters is obvious over the entire water column. Contrasting that with a bottom cold pool of water and an offshore, surface isolated warm pool, reductions in both surface and bottom thermal fronts, with clear dynamical implications, are expected. Therefore, from experiments HeatS to HeatW, theoretically, the bottom thermal front tends to generate a weaker cyclonic circulation around RIS, while the surface thermal front tends to generate a weaker anticyclonic circulation, as shown in Figure 1-17.

Consistent with the above analysis, a decrease in shortwave radiation weakens the bottom and top thermal fronts. Differences between these two experiments (HeatS – HeatW; Figure 1-16b) show a significantly (~ 5 cm/s) reduced northward velocity at the eastern end of the section, and a significantly (~ 5 cm/s) reduced southward velocity at the western end, together comprising a weaker cyclonic circulation in experiment HeatW than in HeatS. The robust changes at the ends of the section are accompanied by weak changes in deeper regions and moving towards the interior. Specially, from 10 Km to 30 Km off Block Island, stronger southward velocities occur in experiment HeatW; from 30 Km to 45 Km, stronger northward velocities occur in HeatW, corresponding to the weak (~ 1 cm/s) anticyclonic circulation difference in Figure 1-16a that is found offshore of the stronger cyclonic difference. According to the above discussion of the schematic shown in Figure 1-17, this feature is attributed to the weakening of the top thermal front in experiment HeatW, and it can explain

why that experiment exhibits a wider cyclonic jet than experiment HeatS. Moving toward the interior of RIS from the inner shelf region, a strong top thermal front in experiment HeatS limits the cyclonic circulation generated by the bottom thermal front where the bottom thermal front is relatively weaker, while in experiment HeatW the decrease of the surface thermal front weakens the resistance to the cyclonic vorticity generated by the bottom thermal front.

Overall, the cyclonic circulation change around the periphery of RIS simulated in the case HeatW is a direct consequence of the reduction of surface heating and its corresponding bottom thermal fronts, while the change in the width of the jet is due to the changes in surface thermal fronts. Therefore, we conclude that bottom thermal fronts are essential to the intensified cyclonic circulation around RIS in summer, with the surface thermal front adjusting the width of the jet. Conducting a similar analysis, we attribute the stronger downshelf current along Long Island to the formation of a stronger bottom thermal front, agreeing with Ullman and Codigas' (2004) study that found an increase in the horizontal density gradient by analysis of historical hydrographic data.

5.3 Geostrophic Balance

To further prove this, we calculate the geostrophic component using the thermal wind equations. Hill et al. (1997, 2008) and Horsburgh (2000) discussed the mechanism of bottom thermohaline gradients based on their observational studies on the northwest European shelf, concluding that the first-order dynamical balance should be geostrophic and the vertical shear in the along-front current should be balanced by the cross-front density gradient through the thermal wind equations:

$$\frac{\partial v}{\partial z} = -\frac{g\partial\rho}{\rho f\partial x};$$

$$\frac{\partial u}{\partial z} = +\frac{g\partial\rho}{\rho f\partial y},$$

where z is taken to be vertically upwards, g is the gravitational acceleration, u is the eastward current and v is the northward current. To compute the absolute velocities from the thermal wind equation we assume a static bottom pool, $u = v = 0$, following Horsburgh (2000), Geostrophic calculations were performed to test whether the observed residual flow pattern is in agreement with the pattern obtained from the baroclinic forcing.

The seasonal mean velocities calculated from the thermal wind equation and the full-physics model results for the case REAL are shown in Figure 1-18. Geostrophic currents in winter are weak, as is expected during this fully stratified season (not shown). In summer, the strong westward jet along the shore of Rhode Island and the weaker offshore current appear in both the thermal wind calculation and the full-physics model results. The flow is mainly geostrophic in summer, except at depth, which is possibly an exaggerated artifact from the simplified thermal wind calculation without considering the bottom Ekman layer and from arbitrarily taking the bottom pool as static – according to Figure 1-15, the bottom currents are weak, within 2 cm/s. Nevertheless, in summer geostrophic currents dominate the cyclonic circulation in RIS.

6. Summary

Dynamical mechanisms associated with the cyclonic periphery current around RIS in summer are discussed based on a series of numerical modeling results and theoretical calculations. The numerical simulations indicate that tidal residual currents and bottom thermohaline gradients are responsible for this summer circulation. The depth-averaged vorticity balance is used to explain the sources of the tidal residual currents in RIS as well as the eddy pair around Montauk Point.

The essential role of tidal rectification in this region is to nonlinearly transfer vorticity from the periodic tidal currents to the mean residual circulation, and the vorticity generated by the tidal currents is dominated by topographic torque. We examined the sources of tidal residual vorticity based on process-oriented experiments and the calculation of the depth-averaged vorticity balance and found that planetary vorticity stretching is the most important process for generating the cyclonic circulation in RIS. The same process is responsible for generating cyclonic eddies inside RIS and anticyclonic eddies to the east of Block Island. On the other hand, the frictional torques due to bottom slope and velocity gradients counterbalance the torque generated by planetary vortex stretching over most regions of RIS.

The formation of tidal eddies around Montauk Point have long been ascribed to 'wall' friction or frictional torque by the topography shoaling toward land (Pingree 1978; Zimmerman 1981; Robinson 1981; Ridderinkhof and Zimmerman 1990) or to nonlinear advection (Pingree 1978; Park and Wang 2000). Our analysis proves the importance of nonlinear vorticity advection in BIS. On the other hand, it shows that frictional torques generate anti-cyclonic eddies over the shallow waters east of

Montauk Point and cyclonic eddies over the deep waters to the south and north of Montauk Point. Moreover, the vorticity balance shows that planetary vortex stretching produces positive vorticity to the east of Montauk Point, but it is less important than the advection term. The major role of the Coriolis force is to trap or turn more waters in BIS, instead of transporting those waters into RIS through the gap between BI and Point Judith.

According to Zimmerman (1981) and Robinson (1981), a strong topographic feature cannot generate significant residual vorticity; only topographic variations that have a length scale within the tidal excursion in the direction of the tidal stream can produce a significant residual vorticity. If the vorticity generating region is considerably larger than the tidal excursion, the oscillating tidal currents are not able to produce residual vorticity, except at the ends of the source region. For a topographic feature within the tidal excursion, the term due to the planetary vorticity is independent of the spatial extent of the topographic feature but depends on the proportional depth change, and therefore is the same for an estuarine or a deeper shelf-sea situation. In contrast, the frictional torque terms (terms e and f in the depth-averaged vorticity equation) are proportional to the velocity of the fluid element as it passes the generating region, and to the length scale of the generating region. Hence for an estuarine and a deeper shelf-sea situation, the terms associated with bottom friction are greatly different.

In a numerical model, to capture the topographic features having the greatest influence in the generation of mean flows, a grid size much smaller than the tidal excursion is necessary (Robinson, 1981). With the tidal excursion larger than 5 km in

most of the studied region, 800 m is an appropriate grid size to solve the tidal residual current with a reasonable computation time. However, at the inner shelf of RIS, there are tidal excursions smaller than 800 m. This means our simulation may underestimate the contributions of the tidal residual current. To better solve the tidal rectification close to Narragansett Bay, a configuration covering RIS with a finer horizontal resolution of 200 m is carried out and its circulation pattern is similar to the 800 m configuration (not shown here), proving 800 m is an appropriate grid size to capture the tidal circulation in RIS with a reasonable computation time.

The impact of bottom thermal fronts in RIS is tested with two experiments; one driven by climatological mean surface heating in August and the other by artificially cutting the short wave radiation in half. Comparisons between the surface circulations and cross-sectional structures of currents and temperature prove the importance of deep thermal fronts for the summer intensification of cyclonic circulation in RIS and the southwestward current along the southern shore of Long Island. Meanwhile, the experiments show that weaker surface heating generates a wider cyclonic jet because of the smaller contribution from a weaker surface thermal front.

Finally, the pool of cold bottom waters offshore of the bottom thermohaline fronts is often static and organic rich, whereas the zones inshore of the fronts are more turbid and rich with phytoplankton. The different water environmental features generate different habitats. The thermohaline fronts thus can provide natural boundaries between different biological communities. Therefore the impact of the construction and destruction of deep thermohaline gradients on the benthic habitat

heterogeneity and primary production are important. According to the observations by Malek et al. (2010) and Nixon et al. (2010), the demersal fish community in RIS was abundant and the phytoplankton density decreased in summer. However, little is known about the exact relationship between the physical structures and biological activities in RIS or elsewhere. This study may provide useful background information for biological oceanographers to study the relationship between the distribution of biological communities and the construction of the deep thermohaline fronts.

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Table

Table 1-1: Tidally averaged transport (unit: m^3/s) along the cruise lines shown in Figure 1-1. Positive represents transport entering the survey region, and negative represents transport exiting the region.

<i>Transect Lines</i>	<i>Seasons</i>	<i>Observations</i>	<i>Model Results</i>
Brenton Reef	Summer	5300+/-300	4334
	Winter	-200+/-250	-220
East Passage	Summer	1100+/-400	776
	Winter	1000+/-500	550
West Passage	Summer	100+/-150	214
	Winter	700+/-100	167
Narragansett Beach	Summer	-5400+/-600	-3123
	Winter	-2400+/-300	-879
South Line	Summer	-100+/-200	-1936
	Winter	300+/-250	332

Figures

Figure 1-1: Bathymetry around RIS and BIS. Place names: Long Island Sound (LIS), Montauk Point (MP), Block Island Sound (BIS), Block Island (BI), Rhode Island Sound (RIS), Narragansett Bay (NB), Buzzards Bay (BB), Vineyard Sound (VS), Martha's Vineyard (MV). The cruise lines at the mouth of Narragansett Bay are represented by black lines: (1) Brenton Reef, (2) East Passage, (3) South Line, (4) West Passage, (5) Narragansett Beach.

Figure 1-2: Schematic of the surface circulation during summer in and around the RIS (Luo et al., 2013).

Figure 1-3: Model configurations. The local-scale ROMS grid (plot every 6 grid points) has a uniform horizontal resolution of 800 m, and the regional-scale ROMS grid (plot every 6 grid points) is uniform with a resolution of 5 Km.

Figure 1-4: Comparisons of tidal sea level variations between model results and observations at the site of PO-S and PO-F (Ullman and Codiga, 2010).

Figure 1-5: Comparison of the depth-averaged velocity fields in May and July between the Model (blue arrows) and observations (red arrows). The field data are from Codiga and Ullman, 2010.

Figure 1-6: Depth-averaged residual currents for the process-oriented experiments of a) M2; b) Flat; c) NoRotate d) NoAdv.

Figure 1-7: Relative vorticity in the case M2 implemented with realistic bathymetry, Coriolis force and quadratic bottom friction formulation with bottom drag coefficient of $C_d = 7.5 \times 10^{-3}$.

Figure 1-8: Relative vorticity in the case NoRotate implemented with realistic bathymetry and quadratic bottom friction formulation with bottom drag coefficient of $C_d = 7.5 \times 10^{-3}$.

Figure 1-9: Residual transport vorticity balance. The relative vorticity is generated by relative vorticity stretching, planetary vorticity stretching, frictional torques associated with bottom slopes perpendicular to tidal currents and velocity shear, transported by advection and dissipated by bottom friction and horizontal diffusions (not shown here).

Figure 1-10: Snapshots of depth-averaged tidal currents in a M₂ cycle with interval of 2 hrs after 1-month simulation. The time evolution is from left to right and top to bottom.

Figure 1-11: An interpretive schematic of the production of tidal residual current in RIS (b) through Coriolis torque. In (a), the flood and ebb stages are represented by heavy solid and dashed arrows, respectively; In (b) Isobaths are shown by two close dotted lines, and the resulting residual circulation is shown by open streamlines. F_c represents Coriolis force.

Figure 1-12: An interpretive schematic of the production of tidal residual current in BIS (b) and RIS (c) through frictional torque (also called slope torque). F_f represents bottom frictional force.

Figure 1-13: Tidal mixing parameter $\log_{10}(H/U^3)$ for M2 tidal currents.

Figure 1-14: Surface currents (a) and cross-sectional structure of along shelf velocity, temperature and salinity (b) in summer (averaged over July, August and September) and winter (averaged over January, February and March) for the experiment driven by monthly climatological atmospheric forcing, rivers, tides and realistic open boundary conditions from regional ROMS (Luo et al., 2013).

Figure 1-15: Time averaged (a) surface circulation and (b) cross-sectional structures of temperature (in °C; black contours) and velocity (in cm/s; color) from experiments (top) HeatS and (bottom) HeatW.

Figure 1-16: Differences of (a) surface circulation and (b) cross-sectional structures of temperature (in °C; black contours) and velocity (in cm/s; color) between the case HeatS and HeatW (HeatS-HeatW).

Figure 1-17: A schematic of the theoretical circulation change from the case HeatS (red) to the case HeatW (green) made by the (left) bottom thermal front and (right) surface thermal front respectively.

Figure 1-18: Geostrophic velocities by thermal wind equation (black lines) and direct model output (colors) across the section athwart RIS in July for the case REAL.

Manuscript 2

Seasonal and Subtidal Variations of the Block Island Sound Estuarine Plume

by

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In preparation for submission to Journal of Physical Oceanography

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Abstract

Buoyant discharge of freshwater from Long Island Sound (LIS) forms a seasonal buoyant plume outside Block Island Sound (BIS) between the coast of Long Island and the denser shelf waters. The plume's seasonal variability and its response to tides, winds and surface heating are investigated through a series of process-oriented experiments using the Regional Ocean Modeling System (ROMS). In winter and spring, the plume is intermediate with a large surface offshore extension detached from the bottom, while in summer, the plume is bottom-advected with most of its width in contact with the bottom and is featured with steep isopycnals. The strong summer insolation together with the weak buoyant discharge and weak winds generates the narrowest and steepest summer plume. In addition, the small changes in tidal currents over the spring-neap cycle cause significant, monthly fluctuations in turbulent mixing and vertical stratification in central BIS, modulating the freshwater discharge which generates episodic fresh water patches that move both downstream along the southern shore of Long Island and toward Rhode Island Sound (RIS) between Block Island and Point Judith. Observational evidence of the detached freshwater patches is discussed in this study.

1. Introduction

Block Island Sound (BIS) is a strait on the inner shelf of the southern New England shelf. It separates Block Island (BI) from the coast of Rhode Island and connects Long Island Sound (LIS) and Rhode Island Sound (RIS) (Figure 2-1), working as the most important passage of freshwater that originates from the Connecticut River.

According to Yankovsky and Chapman (1997), buoyant plumes generated by relatively fresh waters and offshore denser waters are characterized by advective processes that generate three dynamically distinctive plumes: 1) a surface-advected plume that is isolated from any interaction with the bottom; 2) a bottom-advected plume that is controlled by advection in the bottom-boundary layer (Lentz and Helfrich, 2002); and 3) an intermediate plume whose dynamics are a combination of 1) and 2). The Connecticut River, as the purest surface-advected plume that can be found (Garvine, 1974), takes ~70% of the buoyancy (i.e. freshwater) source from LIS and entrains that out of BIS, in the process generating a downshelf buoyant coastal current along the southern shore of Long Island and a bottom-advected plume front to the south of BIS.

The BIS bottom-advected plume has been observed by Kirincich and Hebert (2005) during a 2-day experiment in April 2002, when river discharge is strongest (Figure 2-2). It is accompanied by an along-shelf coastal jet, which is almost linearly sheared with reversed velocities at the bottom and is in thermal wind balance with the mean density field (Yankovsky and Chapman, 1997). In light of the strong dependence of alongshore velocities on the plume and its front, Codiga (2005) used the velocity field

between December 1999 and August 2002 to identify the location of the front and found the plume experiences strong seasonal shifts, involving a shallowest attachment depth in winter, deepest in spring and intermediate in fall, as well as a wider plume width in spring than in winter and fall.

Both the studies by Kirincich and Hebert (2005) and Codiga (2005) have pointed out the importance of the highly variable alongshelf (defined as the direction parallel to the southern coast of Long Island) winds (top panel of Figure 2-2) in the surface offshore extension of the plume due to Ekman dynamics, which is consistent with the available literature (Csanady 1978; Whitney and Garvine, 2006; Moffat and Lentz, 2012). In addition, recent modeling studies by Tilburg (2003) have suggested the importance of winds in the across-shelf direction (the direction normal to Long Island), pointing out that an offshore wind can generate surface offshore transport and an onshore return flow below it, while an onshore wind can generate onshore surface transport and offshore near-bottom transport. From November to March, our study region is subjected to strong offshore and upwelling winds; from July to September, the winds become weaker and there are episodic occurrences of weak downwelling winds in September while the across-shelf direction is dominated by weak onshore winds.

It is very common for a plume to encounter ambient upstream currents. However, most of the available studies have ignored the influence of ambient currents that have been proved important by Chapman and Lentz (1994). In contrast to the assumptions by Csanady (1984) and Wright (1989) that stated upstream flow could move a plume front offshore, Chapman and Lentz (1994) found that this kind of flow limits the offshore spread of a plume and even pushes the front shoreward by imposing an onshore velocity.

Therefore, we believe that the seasonal cyclonic circulation in RIS (the subject of Chapter 1 of this dissertation; also see Kincaid et al., 2003; Luo et al., 2013) not only exchanges waters between RIS and BIS but also limits the offshore spread of the BIS estuarine plume. However, in our study region, this upstream current is closely connected with a bottom thermal gradient, which is associated with surface heating (Luo et al. 2013). Hence, the surface heating may modulate the plume by means of both an upstream current and a bottom thermal gradient.

Tidal currents can significantly affect the river plume dynamics in an intratidal cycle by the tidal asymmetry produced by Strain-induced Periodic Stratification (SIPS; Simpson et al., 1990), and in a subtidal cycle by significant fluctuations in turbulent mixing and stratification over a spring-neap cycle (Peters, 1997; Sharples and Simpson, 1995). Observations in Regions of Freshwater Influenced (ROFIs; Rippeth et al., 2001; Simpson, 1990) and in laboratory experiments (Linden and Simpson, 1986) show that subtidal variations can modulate the transport of buoyancy and form a gravity driven current. Rong and Li (2002) found an eddy-detachment phenomenon at the mouth of Changjiang River and attributed its appearance to the turbulent mixing variation over a spring-neap cycle.

Levine et al. (2009) studied the fronts near the mouths of BIS and LIS, including the BIS estuarine outflow front, the headland front associated with Montauk Point and the Connecticut River plume front, and measured their turbulence and associated mixing at these fronts. However, the spring-neap variability of turbulence and stratification is not clearly defined, neither was the buoyant plume's response to subtidal signals. Moreover, our experiments will show continuous, periodic detachment of an anti-cyclonic eddy, in

the form of an isolated buoyant patch, pinched off from the anticyclonic gyre around Montauk Point when spring-neap tidal variability is included. From surface water salinity observations during spring tides, that is, during 23-25 April, 2001 and 19-20 April, 2002, by O'Donnell and Houk (2009) based on hydrographic surveys, we can notice the appearance of low salinity patches located downstream of the BIS buoyant plume around Montauk Point. But there is no clear evidence that can connect it with the spring-neap variability.

This study will examine the seasonal variability of the plume to the southwest of BIS and its response to seasonally varying winds, river discharge and surface heating by means of seasonal upstream current and deep thermohaline gradients, and we will try to understand the relation between the detached freshwater patches and the spring-neap variations in the tidal turbulent mixing. The numerical experiment design and validation are described in sections 2 and 3. The plume's response to tides is presented in section 4, and its seasonal variation is discussed in section 5. In section 6, we examine the detached fresh water patch generation mechanism. We conclude with a summary in section 7.

2. Design of Numerical Experiments

The numerical model used for this study is the Regional Ocean Modeling System (ROMS), a widely recognized regional and basin-scale ocean model using a high-resolution, free surface, terrain-following coordinate (e.g., Shchepetkin and McWilliams, 1998, 2003, 2005; Haidvogel et al., 2000;). Our configuration, covering the domain of RIS, BIS, LIS and the adjacent inner shelf area, is one-way nested within a larger domain covering the Gulf of Maine/Georges Bank and New England shelf (Figure 2-3; Luo et al., 2013). It has a horizontal resolution varying from 600 m over the RIS and BIS to 1 km

along the boundaries (Figure 2-3) with the number of vertical layers increased from 15 in Luo et al. (2013) to 30 in order to better capture both surface and bottom boundary layers.

A series of experiments are executed to examine the response of the estuarine plume, sourced from the Connecticut River, to tides, wind directions, and surface heating, and their seasonal variability under realistic atmospheric forcings (Table 1). The first experiment, named “Buoy”, isolates buoyant discharge from other forcings. To look at the influence of tides, the second experiment “RivTides” is carried out. It is driven by the five major regional tidal components (M_2 , N_2 , S_2 , O_1 and K_1) as obtained from the Advanced Circulation Model for Oceanic, Coastal and Estuarine Waters (ADCIRC) tidal simulation and a constant Connecticut River point source with an input of $1000 \text{ m}^3/\text{s}$, which represents the climatic mean of all LIS river inputs. This case is also used to analyze the neap-to-spring variations in BIS. The above experiments start from a resting ocean with a uniform salinity of 35 PSU and temperature of 14°C .

The realistic experiment (“Real”) integrates all available, real forcings and is used to understand the seasonal variations of the plume. In addition to tides, the following are used: daily-averaged, climatological atmospheric forcings from the the North American Regional Reanalysis (NARR; Mesinger et al., 2006) from 2004 to 2012; daily-averaged, climatological river discharges for the Taunton River, Blackstone River, Pawtuxet River and Connecticut River as obtained from the U.S. Geological Survey (USGS; <http://waterdata.usgs.gov/nwis>); and open boundary conditions from our ‘regional’ ROMS domain (from Luo et al. 2013). The case Real is driven by climatological daily-averaged data from 2004 and 2012 for 3 years with the first 2 years used for spin-up and the 3rd year results used to analyze the seasonal variability of the plume.

To better represent the salinity input from the Connecticut River we re-scaled the discharges from the USGS to account for the ungauged portions of the watershed by multiplying the total gauged flow by the ratio of total area to gauged area (Chant et al., 2008; Zhang et al., 2010). Because the Connecticut River takes 70% of the riverine input entering LIS, the total discharge into LIS is estimated by dividing the Connecticut River by 0.7 (Figure 2-2). Please note, in this case, the Connecticut River has salinity of 0 PSU and the same temperature with the closest water grid. Moreover, for simplicity, all rivers' input are idealized as point sources at the end of the long channels away from the water points.

In addition to experiment Real, we performed another experiment to examine the winds' contributions to the seasonal variations of BIS estuarine plume, that is, experiment "NoWind" omitting local winds from experiment Real. To test the response of the plume to the four typical winds in the studied region, that is, onshore, offshore, upwelling-favorable and downwelling-favorable winds, a series of idealized experiments driven by buoyant discharge, tides and the corresponding winds are restarted from the 6th month of RivTides for two months (Table 1). These experiments are referred to as WndOn, WndOff, WndUp and WndDown, respectively, in which we set the winds constant as 0.05 N/m^2 . The magnitudes represent the typical wind stresses in the upwelling and offshore directions in winter and fall.

An experiment with halved, constant river discharge from the Connecticut River, referred to as RivTides_500, is compared with RivTides to examine the river discharge's impact on the buoyant plume. For the study of the plume's response to surface heat flux, we performed another pair of experiments by adding constant surface heating from the 6th

month of RivTides and run for another two months. An experiment, “HeatS”, driven by the climatological surface heat flux in August is carried out, and in the other experiment “StratW”, we halve the heat flux to provide weaker surface heating.

3. Model Validation

Luo et al. (2013) have shown that our realistically forced model of the region does an excellent job of simulating tidal residual currents over the whole domain, as well as temperature and salinity at the outer shelf, and Liu et al. (in preparation; Chapter 1 of this thesis) proved its fidelity in RIS. Figure 2-4 shows that our Real experiment simulates the circulation in BIS quite well by comparing its simulated depth-averaged currents in four seasons with observations from the Front-Resolving Observational Network with Telemetry project (Ullman and Codiga, 2004). Specifically, consistent with Codiga’s (2005) observations, in spring, the depth-averaged flow features a robust (~15 cm/s) southwestward current; in summer, our modeling results also agrees well with the limited observations. In winter and fall, the model depth-averaged currents are significantly diminished with more variable directions, which may due to the highly varying winds. The model results show a consistency with the fields in amplitude, but only capture about half of the stations in direction. In light of the highly varying wind directions, the comparison is acceptable. The success in tidal and realistic simulations, as well as our previously published work (Luo et al 2013) indicates that our model successfully represents the essential physical dynamics of RIS and BIS.

To validate the experiment in simulating the river discharge from LIS, we compare velocities across a transect along the Race (marked as section 1 in Figure 2-1) between the case Real and the ADCP observations collected between November 2002

and January 2005 by the Cross Sound Ferry Services of New London, CT (Codiga and Aurin, 2007). Both observations and model results show vigorous water exchange across the channel with surface intensified, fresher current, ~ 15 cm/s, moving out of the estuary along the south of the channel, and bottom intensified, saltier flow moving into the estuary along the northern boundary, as the result of a gravitational circulation. The comparison proves that it is reasonable to treat the rivers in LIS as emanating from the Connecticut River region alone.

To further verify the model in simulating the BIS estuarine plume, we compare the vertical currents in spring with the observations in the FRONT project (Codiga, 2005). As shown in Figure 2-6, consistent with Codiga's (2005) observations, the plume features southwestward currents with robust, surface-intensified along-shelf transport, and with the vertical currents veering clockwise with depth. At the bottom, both the model result and observations show a strong onshore transport even when winds are not strong enough to affect the bottom waters in spring. In brief, the above comparisons indicate our simulation captures the dynamics of the BIS estuarine plume.

4. Plume Response to Tides

The idealized experiment free of impacts of tides and winds, namely Buoy, generates a plume extending broadly to the outer shelf region and to the east of RIS through the gap between Point Judith and Block Island (left panel of Figure 2-7). The plume is steered to the right by the Coriolis force and features a relatively large anticyclonic circulation after it moves out of LIS and BIS. Freshwater coming from the Connecticut River is surface trapped, i.e. without deep signatures of either salinity or

velocity (Figure 2-9). Therefore, the horizontal extension of buoyancy-influenced region must significantly increase to maintain mass conservation, as illustrated in Figure 2-7.

In contrast to the broad surface extension in Buoy, the plume in RivTides is limited to a much smaller area encompassing LIS, BIS, south of BIS and east of RIS with a sluggish horizontal propagation speed, according to the salinity field comparisons (Figure 2-7). After about 4 months' simulation, the structure of the buoyant plume reaches a steady state with a confined offshore extension.

Figure 2-8 depicts the pattern of the surface buoyant plume in the 8th month, showing Lagrangian trajectories of surface drifters that originated in the mouth of LIS released at the first day of the 8th month. Although the trajectories are influenced by the stage of tides when drifters are released and a large uncertainty exists for such few particles (Ullman et al., 2006), it is still useful to predict the water dispersion nonetheless. The patterns of the trajectories are in agreement with the trajectory prediction by Ullman et al. (2006). The simulation shows that the drifters in the southern part of the channel move southward through BIS and then move downshelf or offshore exiting the study domain, and the particles originating in the northern part move westward into RIS and then are entrained back into BIS.

The difference between Buoy and RivTides derives from the deep penetration of freshwater in RivTides, and is consistent with the available literatures, for example Simpson (1997), who emphasized the importance of tidal mixing in the plume spread. At the same time, the case RivTides generates a constant upstream flow, ~ 3 cm/s, from RIS imposed on the plume to the southwest of BIS resulting from the topographically-induced tidal rectification (Liu et al., in preparation, and Chapter 1 of this thesis). This current

works to turn the trajectories of the particles into RIS back to BIS. Therefore, waters entering RIS through the gap between Block Island and Point Judith are isolated from RIS waters. According to Chapman and Lentz's (1994) study, because of the limitation by the upstream current, we infer that the plume in RivTides should be narrower and have a larger horizontal density gradient than the theoretical estimates of Yankovsky and Chapman (1997). In the following discussions the effects of tides will be implicitly included.

The middle panel of Figure 2-9 shows the cross-shelf sections of the alongshelf and cross-shelf velocities and salinity along the section outside BIS, normal to the southern shore of Long Island (section 4 in Figure 2-1) after 8 months' evolution. The total width of the plume is $W_p = W_b + W_s$, and is composed of the part in direct contact with the bottom (W_b) and the part away from the bottom, from the bottom offshore edge of the plume to its surface offshore edge (W_s). Since $W_b < W_s$, the plume generated by RivTides is classified as 'intermediate', with isopycnals sloping from the bottom at about 22 km offshore of the coast to the surface at 20 km further offshore. The trapping depth, also called the attachment depth, at which the bottom-advected plume becomes trapped (Yankovsky and Chapman, 1997) with the onshore and offshore velocities reaching a balance, is within 20 m for the transect chosen, much smaller than the estimation by Kirincich and Hebert (2005) and Codiga (2005) at a further upstream region, around the mouth of BIS.

The alongshelf flow is surface-intensified with almost linear vertical shear, reversing within several meters of the bottom at the foot of the front. It is geostrophic, by comparison of the full physics model output with a thermal wind calculation (Figure 2-

10) and by the diagnostic analysis of the across-shelf momentum balance (black lines in Figure 2-11) - during the thermal wind calculation, we approximately set the bottom, weak velocity within 1 cm/s as 0, and in the diagnostic analysis, we cast the momentum equations into a right-handed coordinate system with x directed upshelf along the shore of Long Island and y directed onshore. Meanwhile the offshore flow responsible for the offshore spread of the inner-shelf fresher waters penetrates to the depth where the alongshelf velocity reverses. Such deep onshore flow has been observed in the southwest of BIS by the moored profiling current-meter records in spring when wind effects are excluded (Codiga, 2005; Figure 2-6). Diagnostic analysis of the along-shelf momentum balance in Figure 2-11 verifies Yankovsky and Chapmans' (1997) hypothesis that the deep onshore transport is due to bottom Ekman dynamics with a balance reached by the pressure gradient, Coriolis and vertical viscosity terms.

5. Seasonal Variability of the Plume

We defines the average from January to March as winter, April to June as spring, July to September as summer and October to December as fall. The comparison between the simulated depth-averaged currents and observations from the FRONT project (Ullman and Codiga, 2004) shows a strong seasonal cycle of circulation. Also, the surface salinity (Figure 2-12) and the transect along the section normal to Long Island (Figure 2-13) demonstrates a strong seasonal variability of the BIS estuarine plume generated by the freshwater emanating from LIS.

BIS is fresher in winter and spring than in summer and fall, with the widest and strongest buoyant plume in spring when river discharge peaks and the narrowest and weakest plume in summer when river discharge is smallest (Figure 2-13). Compared with

spring, freshwater in winter is limited in its downshelf penetration and is a saltier plume, though their offshore spreading is comparable. Moreover, the drifters in winter are swiftly diverted offshore, and the drifters in fall are steered more offshore than in spring and summer, which should be due to the strong and prevailing upwelling winds.

The transect normal to Long Island reveals an intermediate plume in spring with the bottom offshore edge (32.6 PSU) around 18 km offshore, and is accompanied by a robust, surface-intensified downshelf transport, as well as a strong surface offshore current across the salinity front. In the across-shelf direction, at depth, there are currents opposite to the shallower transport even when winds are negligible, which is associated with the bottom Ekman layer. Compared with the other seasons, during spring the waters on the landside of the front are freshest; the front has the sharpest gradient and widest frontal width with a slope angle around 0.001. The slope for the isohalines at the offshore edge of the plume is even smaller. On the landside of the front, there is a strong onshore current over the water column, which may be a part of a headland eddy around BIS (see Chapter 1 for a dynamical analysis of this feature).

Though there are not enough observations in summer to delineate the buoyant plume, the fact that our modeling result agrees well with the limited current observation gives us confidence in presenting the summer features. During this season, a robust downshelf jet larger than 16 cm/s moves through the salinity front. The summer plume is bottom-advected with a narrower plume width both at the bottom (~12 km) and at the surface (~22 km), and steeper frontal slope (~0.002 for the isohaline of 32 PSU, double the slope in spring) than the other seasons. Also, its bottom upshelf layer almost

disappears and the onshore deep transport is weak and confined to a narrower region. Similar to spring, on the landside of the front, there is onshore transport.

In winter and fall, the plume is also intermediate. As described by Codiga (2005), the alongshelf transport is significantly weakened and moved offshore with the shallowest penetration depth of the shallow transport, which is especially obvious in winter's result; in winter, the deep upshelf jet is much thicker than the other seasons. The across-shelf flow produces a surface offshore transport, even on the landside of the front in winter, corresponding to a broad surface extension with plume widths about 15 km in fall and 20 km in winter, generating slope angles smaller than summer but larger than spring with a value about 0.015 in winter and 0.018 in fall for isohaline of 32 PSU. Another important feature is the obviously strengthened deep onshore transport: the deep onshore transport spans across the bottom of the entire transect and extends even more offshore.

5.1 Response to Winds

The experiment NoWind, which omits the effects of local winds, produces a bottom advected plume in winter and fall with a strengthened downshelf transport and a weakened offshore transport across the front in the surface layers, as well as a weakened and thinner deep upshelf transport and onshore transport. Moreover, the center of the surface downshelf transport is located more towards the inner shelf than in the experiment Real.

In spring, the plume width in NoWind is narrower than in Real. Taking the isohaline of 32 PSU as an example, its surface outcropping position moves from 35 km

offshore to 30 km offshore with its contact with the bottom kept intact. Another important feature in the NoWind experiment is that, especially in spring, though the pattern of the front keeps intact compared with the case Real, at the inner shelf the water becomes fresher, creating a larger salinity gradient in Real. This may be due to the change of freshwater delivery out of BIS caused by omission of the winds.

Freshwater delivery is quantified by a freshwater flux that is defined as:

$$F = \int_{l_s}^{l_e} \int_{-h}^{\zeta} \left(\frac{S_a - S}{S_a} u \right) dz dl \quad (1)$$

where S_a represents ambient salinity, S is salinity and u represents velocity across the section. The flux is integrated over the whole depth from the bottom ($-h$) to the sea-surface displacement (ζ) over the length of the domain starting with l_s and ending with l_e . Table 2 lists the freshwater delivery out of LIS and BIS across sections 2 and 3 in Figure 2-1 for the cases Real and NoWind during different seasons. In Real, freshwater flux through BIS shows significant seasonal variations with a maximum value (about 900 m^3/s) when the river discharge peaks. In the NoWind experiment, the seasonal variability still exists, however the absence of wind decreases the freshwater flux out of BIS in winter and fall, while increasing the values in spring and summer.

To further examine the roles of wind, we analyze the four idealized experiments driven by winds in different directions: WndOn, WndOff, WndUp and WndDown. Figure 2-15 shows the responses of surface salinity after 3 days of wind forcing and the Lagrangian trajectories of the surface drifters during the initial 5 days, in which both the buoyant plume and the trajectories vary with wind direction. In every case, upon leaving the Connecticut River, most of freshwater is stored in LIS, while a smaller fraction is

transported out of LIS and mixed with the outer shelf saline waters. As shown by surface salinity and freshwater flux, an upwelling-favorable wind is most effective in spreading the freshwater offshore and a downwelling-favorable wind is most effective in squeezing the plume against the Long Island coast.

Table 3 indicates that the presence of upwelling-favorable and offshore winds transport more freshwater out of BIS than the case free of winds, though the delivery out of LIS is smaller in WndOff than RivTides, while the onshore and downwelling-favorable winds suppress the freshwater delivery. After exiting LIS, the upwelling wind rapidly spreads the buoyant plume eastward to the broader outer shelf region through BIS at a rate of $785 \text{ m}^3/\text{s}$, and through the gap between Point Judith and Block Island at a rate of $256 \text{ m}^3/\text{s}$. By contrast, the offshore wind constrains most of the freshwater from LIS to the outer shelf region directly through BIS – the freshwater flux through BIS is $608 \text{ m}^3/\text{s}$ when leaving LIS at the rate of $614 \text{ m}^3/\text{s}$.

The application of onshore and downwelling-favorable winds retain more freshwater in LIS by decreasing the freshwater flux leaving LIS to $530 \text{ m}^3/\text{s}$ in WndOn and $331 \text{ m}^3/\text{s}$ in WndDown. Even though the flux in WndDown is small, its wind is much more effective in delivering freshwater to the BIS estuarine plume, while the onshore wind delivers more than half of the freshwater going eastward into RIS through the gap between Point Judith and Block Island. Note that most of the freshwater going into RIS is entrained back to BIS through south of Block Island, which explains why the plume appears to extend further offshore under onshore winds than offshore winds.

The drifter trajectories shown in Figure 2-15 agree with the freshwater flux in Table 3, from which we find that the upwelling-favorable wind rapidly spreads the

drifters released at the mouth of LIS eastward through both RIS and BIS, while the offshore wind drives all the drifters, even those closest to the northern shore of BIS, southward. Moreover, the onshore wind pushes the drifters eastward into RIS, and in the experiment with downwelling-favorable wind, drifters move around the mouth of LIS with little chance of leaving LIS.

The above analysis can explain the freshwater flux change from Real to NoWind. During fall and winter, the study region experiences prevailing upwelling-favorable and offshore winds, both of which strengthens the freshwater advection into the buoyant plume. Therefore, omitting local winds results in a weaker freshwater flux leaving BIS. In spring, the upwelling-favorable wind and onshore wind exert opposite influences on the freshwater flux; unlike an upwelling wind, an onshore wind suppresses the freshwater delivery, which seems more important to the buoyancy transport in spring by holding more freshwater inside the estuaries (Table 2). The winds in summer act similarly, though the amplitude is much smaller.

The winds' effect on freshwater dispersion are accompanied by the changes in the vertical current structure (Figures 2-16). In the alongshelf direction, an onshore wind reduces the downshelf velocities at the surface via Ekman dynamics; in the across-shelf direction, comparing WndOn with RivTides, we find an onshore wind generating a surface onshore current (~ 2 cm/s) and a return flow directly below it, responsible for the sharp halocline around 10 m depth. Such a structure is in agreement with Tilburg's (2003) study that found a significant surface transport and a return flow below it though the depth-integrated transport is feeble. Similarly, an offshore wind produces an offshore transport at the surface and a deeper returning, onshore transport below it. At the same

time, the offshore wind triggers stronger alongshelf velocities via strong surface and bottom Ekman dynamics.

Except in a thin layer around 10 m and offshore of 10 km, an upwelling-favorable wind drastically alters the alongshelf flow by reversing the downshelf currents over the water column. In the cross-shelf direction, an upwelling wind drives the waters shallower than 10 m offshore and activates a stronger bottom return current bringing about an intrusion of offshore saline waters, which detaches the core of the buoyant plume from the coast and diverts the downshelf transport of freshwater offshore. Conversely, a downwelling-favorable wind creates a strong onshore transport in shallower layers and a deep offshore flow at the bottom, at the same time accelerating the downshelf transport and squeezing the front into a much narrower band.

Accordingly, the variations in the vertical structures of the buoyant plume in Real can be explained as follows. In winter and fall, the upwelling-favorable wind is stronger than the offshore wind, which makes the effects by the upwelling winds more prominent by decelerating the downshelf current and generating a stronger offshore surface transport and a stronger deep onshore transport. In spring, the upwelling-favorable winds and the relatively weaker onshore winds have the same impact on the downshelf transport. So they should work together to slow down the downshelf flow. However, in the across-shelf direction, they work against each other, that is to say, they counterbalance each other to only moderately push the surface freshwater offshore. Winds in summer are weak and their influence to the plume structure is negligible.

In summary, the upwelling-favorable wind is the driving force behind the following characteristics in winter and fall: 1) the broader surface extension of the plume,

2) the shallower downshelf velocity penetration, 3) the stronger deep up-coast flow and 4) the stronger deep onshore flow. In addition, wind also works to modestly broaden the surface horizontal extension in spring.

5.2 Response to River Discharge

Besides wind effects, buoyant discharge variations can result in changes in the plume and its front (Chapman and Lentz, 1994). To examine this, we configure two idealized experiments driven by constant river and tides only: RivTides with a constant river discharge of $1000 \text{ m}^3/\text{s}$ and RivTides_500 with half of that river discharge ($500 \text{ m}^3/\text{s}$). The resulting buoyant plumes are shown in the middle and bottom panels of Figure 2-9 whereby we find the stronger river input broadens the plume's width by more than 10 km at the surface and increases the strength of the front. The doubled river discharge almost doubles the salinity difference between the seaside edge and landside edge, which is ~ 1 PSU in RivTides_500 and 2 PSU in RivTides. Therefore, in light of the pronounced seasonal variation in LIS river input, we can infer that the increase in river discharge from winter to spring as shown in Figure 2-2 contributes significantly to the sharper and broader plume in spring, especially when the river discharge peaks in April, and the abrupt decrease from spring to summer should be responsible for the weaker and narrower plume in summer, when discharge subsides to the bottom.

On the other hand, the significant change in river input only slightly changes the attachment depth of the front and the offshore extension of the plume and has a limited influence (within 1 cm/s) on the cross-shelf and alongshelf velocities. However, in the NoWind experiment we noticed much more obvious changes in the velocities and plume's offshore extension. This fact together with Chapman and Lentz's (1994) prompt

us to study the effects of other factors, that is, the upstream current from RIS and the deep thermal fronts, which are both related to the significant increase in surface heating in summer (Liu et al, in preparation and Chapter 1 of this thesis).

5.3 Plume Response to Surface Heating

Chapman and Lentz (1994) paid attention to the upstream flow's impacts on the evolution of the buoyant plume, which is highly seasonally variable in BIS and worthy of note. Our studies (Liu et al., in preparation) have found that the increase in surface heating forms a stronger deep thermal front and a stronger cyclonic circulation around RIS and, therefore, a stronger upstream current imposed on the plume front. In addition, the comparison of surface salinity indicates that the increase in surface heating pushes the surface plume onshore (Figure 2-17), which is resulted from the increase of the upstream current from RIS according to the theory by Chapman and Lentz (1994).

In the vertical, compared with the nonstratified case RivTides, the increase of surface heating in HeatW and HeatS significantly increases the surface-intensified downshelf current with a narrower and weaker upshelf current, at the same time deepening and strengthening the surface-intensified offshore transport. At the surface the offshore transport increases from ~ 1 cm/s to ~ 3 cm/s from RivTides to HeatW, while in the bottom layers, the change in the onshore deep velocities is much smaller (slightly weaker and limited to a thinner layer).

Figure 2-11 shows the momentum balance in the alongshelf and across-shelf directions for the experiments RivTides, HeatW and HeatS. The momentum balance in

the alongshelf direction is achieved by pressure gradient, Coriolis force and alongshelf momentum advection, while the vertical viscosity plays an increasingly important role at depth. The analysis shows a significant increase in the alongshelf pressure gradient from RivTides to HeatW, especially in the shallower layers, and is responsible for the increase in the cross-shelf velocities. Meanwhile, note that there is an increase in u-momentum advection in the alongshelf direction at depths greater than 15 m. The analysis of the momentum balance in the across-shelf direction shows that the alongshelf velocity is geostrophic and the increase in downshelf current is due to the increase in the across-shelf pressure gradient.

As discussed above, the changes in pressure gradients $(-\frac{1}{\rho_0} \frac{\partial P}{\partial x}, -\frac{1}{\rho_0} \frac{\partial P}{\partial y})$ dominate the response of currents and plume to surface heating. Following Csanady (1979) and Ullman and Codiga (2004), the pressure gradients can be decomposed into the part arising from density gradients (baroclinic component) and the part from sea surface elevation gradients (barotropic component):

$$-\frac{1}{\rho} \frac{\partial P}{\partial x} = g \frac{\partial}{\partial x} \int_0^z \epsilon \, dz' - g \frac{\partial \zeta}{\partial x} \quad (2)$$

$$-\frac{1}{\rho} \frac{\partial P}{\partial y} = g \frac{\partial}{\partial y} \int_0^z \epsilon \, dz' - g \frac{\partial \zeta}{\partial y} \quad (3)$$

where P is the pressure, ζ is the free surface elevation and ρ is the density relating to a reference density anomaly (ρ_0) and a dimensionless density anomaly (ϵ), $\rho = \rho_0[1 + \epsilon(x, y, z, t)]$. The contributions from the baroclinic and barotropic parts are shown in Figure 2-19 as solid and dashed lines respectively.

Consistent with Chapman and Lentz's study (1994), a stronger upstream current from RIS imposes a stronger downshelf pressure gradient arising from the sea level variation. However, different from their investigation, in BIS the pressure gradient arising from baroclinic part is significant, offsetting the effects by changes in sea surface height and creating a stronger upshelf pressure gradient and therefore stronger offshore velocities. Further examining the respective effects of temperature and salinity on density by keeping one of the hydrographic terms constant (not shown here), we find the density gradient in the alongshelf direction mainly arising from salinity.

Similar analysis in the across-shelf pressure gradient also finds opposite roles played by the barotropic and baroclinic components; the sea level variation produces a stronger onshore pressure gradient, while the density gradient offsets the effect and contributes to a stronger offshore pressure gradient that is responsible for the stronger downshelf current in the experiments with stronger surface heating. Further examination proves the density gradient is arising from temperature, more specifically, from the formation of a deep thermal front. Thus surface heating controls the BIS plume by way of an upstream current and a deep thermal front, with the effect by the deep thermal front dominated in the across-shelf velocity.

The changes in vertical velocity structures from RivTides to HeatS are accompanied by a steeper and narrower plume over the whole depth – the bottom edge of the plume (isohaline of 34.8 PSU) shoals from 24 km offshore in RivTides to 18 km in HeatW to 14 km in HeatS. This can be explained by the increase in the alongshelf freshwater transport ($u \frac{\partial s}{\partial x}$) because of the increasing u , which suppresses the buoyancy's offshore spreading. According to the above analysis, we can conclude that, in addition to

the seasonally varying wind and buoyancy discharge, surface heating in spring and summer plays an important role in the buoyant plume by steepening the isohalines and narrowing the plume, in agreement with Ullman and Codigas' (2004) study stressing the importance of horizontal density gradients through analysis of historical hydrographic data.

6. Generation of Periodic Detached Freshwater Patches

The simulation with spring-neap tidal variability (RivTides) shows continuous, periodic anti-cyclonic eddies, in the form of an isolated buoyant patch, detached from the BIS estuarine outflow front, as shown in Figure 2-20. Before detachment, BIS features anticyclonic currents around Montauk Point and a buoyant bulge structure. Around day 192, when the strong spring tide is about to happen, an isolated freshwater patch is formed (defined by closed salinity contours) and surrounded by anticyclonic motions (i.e., negative relative vorticity), separated from the headland eddy by a narrow band of positive relative vorticity. After detachment, the isolated patch is transported downshelf along the southern shore of Long Island and dissipates along its propagation pathway. During the propagation, the anticyclonic eddy with negative relative vorticity remains.

The surface salinity delineates another isolated, freshwater patch that is shed from the freshwater plume in central BIS and moves into RIS through the gap between Block Island and Point Judith. Both patches have been observed by the limited salinity observations by O'Donnell and Houk (2009) on spring tides during 23-25 April, 2001 and 24-26 September, 2001 (Figure 2-21). During 23-25 April, 2001, a pool of freshwater with salinity of 29.8 PSU is observed downstream of the headland eddy around Montauk

Point. Meanwhile, the survey captured the fresh salinity pool, with salinity of 30 PSU, between Block Island and Point Judith.

The detachment in experiment RivTides has a monthly period, corresponding to the period of large spring tide. The experiment driven by the river and M2 and S2 also generates such detached patches with a regular fortnightly period, consistent with the subtidal period of M2 and S2, and a smaller length scale. Such eddies disappear in experiments that omit either river discharge or spring-neap variability, i.e., for a case driven by the Connecticut River and only the M2 tides, there is no eddy detachment. For our study area the detached eddies need both a continuous buoyancy input and the spring-neap variability of tides. Rong and Li (2012) attributed a similar eddy found at the mouth of Changjiang River to turbulent mixing variations over a spring-neap cycle. In the following, we will use the case RivTides to understand how tides affect the turbulent mixing and stratification in BIS, and therefore trigger the BIS buoyant plume to generate its periodic shedding.

Before proceeding to the discussion of the spring-neap variability of turbulent mixing caused by tides, we need to first explain a phenomenon that is called the Strain-Induced Periodic Stratification (SIPS) by Simpson et al. (1990), which arises from the flood-ebb asymmetry over a tidal cycle. Over a tidal cycle, the stratifying influence that results from the freshwater input from LIS competes with the de-stratifying influence by tidal stirring, producing significant flood-ebb fluctuations with maximum stratification during the ebb when freshwater in the upper layers is moved seaward over saltier water in the deeper layers, and less stratified conditions during the flood when freshwater in the upper layers retreats (Simpson et al., 1990; Rippeth et al., 2001). The periodic switching

between stratified and well-mixed conditions over a tidal cycle is referred to as strain-induced periodic stratification (SIPS) by Simpson et al. (1990).

Whitney et al. (2012) found that, in some regions of LIS, the interaction between straining and stirring would shift the maximum stratification compared with a straining-only case. This feature has been captured by our model results, but it is beyond the scope of this manuscript. However, the switching between stratified and less-stratified (or mixed) conditions remains. In addition to the flood-ebb fluctuations, the horizontal gradients interact with tidal stirring over the spring-neap cycle to control the dispersion of freshwater in BIS according to the studies by Peters (1997) in the Hudson River off Manhattan and by Sharples and Simpson (1995) in Liverpool Bay.

Levine et al. (2009) studied the fronts near the mouths of BIS and LIS, namely, the BIS estuarine outflow front, the headland front associated with Montauk Point with and without sandwaves, and the Connecticut River plume front. Different types of fronts respond differently to tidal straining and turbulent mixing that vary over a spring-neap cycle. To analyze the spatial and temporal variations and understand the interaction between turbulence, stratification, and shear in BIS, we selected three stations; station A close to the Connecticut River plume front, station B within the headland front, and station C in the BIS estuarine plume front (all marked in Figure 2-20).

To describe the competition between the destratifying process of turbulent mixing and the stratifying process of tidal straining, a dimensionless number, referred to as the Simpson number (Si), has been used (Simpson et al., 1990; Stacey et al., 2001).

$$Si = \frac{(\partial_x b) H^2}{U_*^2} \quad (4)$$

where $\partial_x b$ is the depth-averaged tidal mean horizontal gradient of buoyancy b ($b = -g \frac{\rho - \rho_0}{\rho_0}$ with ρ_0 representing the reference density; g , gravitational acceleration; ρ , potential density), H represents the water depth, and U_* is the friction velocity scale. Since a quadratic bottom drag formulation is used, we approximately have $U_* = \sqrt{C_D} U$ with C_D representing the bottom drag coefficient and U the root mean square tidal velocity. The threshold between the well-mixed and SIPS stages is $Si = 0.088$ while the threshold between the SIPS and permanently stratified stages is $Si = 0.84$ (Becherer et al., 2011).

The level of stratification is represented as the salinity difference between the bottom and surface waters, ΔS , and the spring-neap variability of turbulent mixing in BIS can be represented by vertical turbulent buoyant fluxes B , which is estimated using $B = K_\rho g \frac{\partial \rho}{\partial z}$, where K_ρ is the vertical salinity diffusivity, g is gravitational acceleration, and ρ is potential density. Figure 2-22 shows the time series of the Simpson Number, and Figures 2-23 to 2-25 show the variations of stratification and vertical buoyancy flux at these stations.

At the Connecticut River plume front station A, the Simpson number is larger than 0.84 most of the time except during the stronger spring tides. Correspondingly, tidal stirring produces spring-to-neap variations in stratification and vertical buoyancy flux; Station A remains periodically stratified (SIPS) with high tidally-averaged buoyant flux during the stronger spring tide, while permanently stratified with different stratifications (ΔS varying around 4PSU) over a tidal cycle and significantly diminished tidally-averaged buoyancy flux during the neap and weaker spring tides. The monthly variation

agrees with the idea by Li and Zhong (2009) that the stratification build-up depends on both the magnitude of the mixing coefficient and its time history.

Compared with station A, the headland front station B, with a Simpson number smaller than 0.088 everywhere, experiences much smaller spring-neap variation, because the strong tidal turbulence is always well-mixing the water column, making tidal rectification effects typically dominate over that of both wind stress and buoyancy forcing. Hence during a spring-neap tide, within the headland front, the salinity stratification remains small (i.e., the flow remains well-mixed) except for a short period during the stronger neap, and the buoyancy flux remains constant. Within the estuarine outflow front station C, where the Simpson number is always larger than 0.84, the water column should be permanently stratified over the entire spring-neap cycle, unlike stations A and B that transform between homogeneous and stratified conditions. Figure 2-25 shows a modest, with changes of salinity within 1 PSU, spring-neap variability of salinity stratification, larger than station B but smaller than station A. At the same time, the variation of turbulent buoyancy flux is small, which proves the variation of stratification is caused by the variation of freshwater transport out of BIS.

The periodic transition from SIPS to permanent stratification at the central and north BIS dramatically modulates freshwater delivery out of BIS. Figure 2-26 shows the spring-neap cycle in freshwater transport with the maximum transport around neap tides and minimum around spring, even though a constant river discharge is supplied at the Connecticut River. The freshwater delivery variation across BIS resulting from the spring-neap variability of turbulent mixing works equivalently with the variable discharge from a river that is described by Yankovsky et al. (2001) as generating the

detached anticyclonic, freshwater patch when freshwater transport subsides after reaching its peak value. To examine the momentum balances, we analyze the surface diagnostic terms in the alongshelf direction (Xi-direction) along section E and in the across-shelf direction (Eta-direction) along section X. These sections cross the detached patch south of Montauk Point (Figure 2-20).

The baroclinic Rossby radius ($R_d = \frac{\sqrt{g'h_0}}{f}$, where $g' = g|\Delta\rho|/\rho_0$) of the inflow is about 6 km. Within the buoyant plume, offshore of the baroclinic Rossby radius ($y \geq R_d$), the surface flow is in geostrophic balance with little contributions made by the advection terms. Correspondingly, the downshelf velocity is driven by the cross-shelf pressure gradient. Moving shoreward along section E, the Coriolis force term becomes smaller. By contrast, the horizontal advection of U-momentum in Xi-direction ($-u \frac{\partial u}{\partial x}$) grows. Within the Rossby radius ($y \leq R_d$), a balance is established between the pressure gradient and horizontal advection, responsible for an ageostrophic flow that is much weaker than the downshelf geostrophic flow further offshore or even reverses upshelf, opposite to the geostrophic transport. Such ageostrophic upshelf transport within the Rossby radius can be found at day 198 in Figure 2-20.

Along section X, away from its east end, where strong salinity and velocity advection occur, the surface flow is in geostrophic balance as shown in the bottom panel of Figure 2-27. As for the part close to BIS, geostrophic terms are still important, but at the same time horizontal advection, especially advection in the Eta-direction, becomes important in the balance. Thus the above analysis paints the following picture of the eddy generation processes southeast of BIS. Freshwater transported out of LIS through BIS

retreats from neap to spring, isolating the freshwater patch flowing out of BIS before the retreat to the southeast of Montauk Point. Then this patch generates geostrophic anticyclonic currents around it where the offshore distance is larger than the baroclinic Rossby radius and the ageostrophic velocity within the baroclinic Rossby radius. When the ageostrophic velocity turns upshelf, a closed anticyclonic circulation will be formed around the detached freshwater patch. The eddy then propagates downshelf along the coast, during which it dissipates quickly.

The model produces a width of the freshwater patch in the across-shelf direction during the detachment of about 25 km. According to the generation mechanism, it is consistent with the width of the buoyant plume at the peak of freshwater transport from BIS. During the propagation, the patch shrinks, which may be a result of lateral mixing or diffusion. From day 192 to day 198, the core of the detached pool moves about 17 km downshelf, corresponding to a velocity of ~ 3 cm/s. In an experiment with doubled river discharge, the core of the patch becomes obviously fresher, but there is no significant change to its propagation speed and its size. This proves the length scale and propagation are not sensitive to the reduced velocity. The along coast propagation speed of the detached patch is likely a combination of advective and coastally-trapped wave processes; this component of the process needs further study.

7. Summary

A seasonally varying buoyant plume has been observed southwest of BIS. This study uses a numerical model and theoretical diagnostics to investigate its seasonal variability and its response to tides, winds and surface heating. Idealized model experiments indicate that tidal mixing is required in the formation of the bottom-advected plume southwest of BIS by isolating the freshwater coming out of LIS to a deeper depth and trapping the buoyant plume in a confined region.

The effects of winds on the buoyant plume are examined by two process-oriented experiments and a series of idealized experiments driven by the typical winds – onshore, offshore, upwelling-favorable, and downwelling-favorable winds. Results of our realistic experiment show an intermediate plume in winter and spring, when the upwelling winds dominate and spread the surface freshwaters offshore, and a bottom-advected plume in summer with the steepest front. Analysis of the idealized experiments reveals that winter-prevailing upwelling and offshore winds are responsible for the broad surface extension of the plume in winter. In summer, this region is subjected to weak onshore and downwelling-favorable winds, which resists the delivery of freshwater out of BIS and traps the plume closer to the coast.

In addition, the seasonal variability in river discharge, mainly from the Connecticut River, plays an important role in the plume's width and strength. When river discharge peaks in spring, the plume is widest and has a very small frontal slope about 0.001. The decrease in river discharge from spring to summer contributes to the narrower and weaker plume. However, the river discharge's impact on the offshore edge at the bottom and the alongshelf and across-shelf velocities across the salinity front is limited.

This study demonstrates that the seasonal variation of surface heating is a competitive contributor to the seasonal plume. According to the idealized experiments driven by different extent of surface heating, we found that the increase of surface heating eventually produces a plume with narrower offshore extension both at the surface and bottom. Meanwhile, both the offshore transport and downshelf transport across the front becomes stronger, which is mainly due to variations in density gradients. Thus in spring, the growing surface heating increases the upstream current from RIS and forms a bottom thermal front, which tends to compete with the spring discharge peak by shoaling the bottom offshore edge of the plume and limiting the surface spreading of buoyancy. In summer, the narrowest and shallowest plume occurs because of a combination of several physical processes. First, the smallest river discharge limits the freshwater transport into the plume. Second, the appearance of weak onshore and downwelling favorable winds tend to push the surface plume onshore. Finally, the strong surface heating pushes the bottom offshore edge of the plume to an even shallower depth and narrower band.

Our experiments find an important feature in BIS that has been observed by O'Donnell and Houk (2009) – the periodic detached freshwater patches shedding off from the buoyant plumes northeast and southwest of BIS. These patches are formed during the strong spring tides and are associated with the subtidal (spring-neap) variability in tidal turbulent mixing, especially in regions within the Connecticut River plume. Analysis shows the spring-neap variability in turbulent mixing generates subtidal variation in the freshwater flux across BIS, working equivalently with the variable discharge from a river. When the freshwater flux becomes weakest, anticyclonic vorticity is generated around the isolated patch south of BIS and propagates downshelf. This

manuscript supplements the studies by Yankovsky et al. (2001) that focused on the detachment of anticyclonic eddies on a surface-advected buoyant plume.

Acknowledgement

We thank Dan Codiga and Dave Ullman for graciously shared their observational data with us. We acknowledge the international modeling groups for providing their data for analysis. Q. Liu would like to thank Dave Ullman, Chris Kincaid and Jason Dahl for their insightful comments on her work.

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Tables

Table 2-1: List of Experiments with ROMS.

Run	Tides	Wind	River Discharge	Surface Heating
<i>Buoy</i>	No	No	1000	No
<i>RivTides</i>	Yes	No	1000	No
<i>RivTides_500</i>	Yes	No	500	No
<i>Real</i>	Yes	Daily Data from NARR	Daily Data from USGS	Daily Data from NARR
<i>NoWind</i>	Yes	No	Daily Data from USGS	Daily Data from NARR
<i>WndOn</i>	Yes	Onshore	1000	No
<i>WndOff</i>	Yes	Offshore	1000	No
<i>WndUp</i>	Yes	Upwelling	1000	No
<i>WndDown</i>	Yes	Downwelling	1000	No
<i>HeatS</i>	Yes	No	1000	Aug Surface Heating
<i>HeatW</i>	Yes	No	1000	0.5×Aug Surface Heating

Table 2-2: Freshwater flux through BIS for the cases Real and NoWind in different seasons

Run	Winter	Spring	Summer	Fall
<i>Real</i>	360	906	471	340
<i>NoWind</i>	317	995	509	298

Table 2-3: Freshwater flux across the sections in Figure 2-1.

Freshwater Flux (m^3/s)	WndOn	WndOff	WndUp	WndDown	NoWind (RivTides)
<i>Section 1 (Out of LIS)</i>	530	614	1041	331	630
<i>Section 2 (Out of BIS)</i>	210	608	785	371	496

Figures

Figure 2-1: Bathymetry and geographic features around the Rhode Island Sound. Place names: Connecticut River (CR), Long Island Sound (LIS), Rhode Island Sound (RIS), Block Island Sound (BIS), Narragansett Bay (NB), Buzzards Bay (BB), Vineyard Sound (VS). The thick red line is the section used to describe the buoyant plume to the southwest of BIS, and the thick black lines are the sections referenced in Table 2.

Figure 2-2: Wind and Long Island Sound river input. (Top panel) daily winds and the corresponding alongshelf (blue) and across-shelf (green) components from NARR averaged from 2003 to 2012. (Bottom panel) Daily Long Island Sound river input, estimated from the Connecticut River discharge by dividing the Connecticut River by 0.7, from USGS.

Figure 2-3: Model configurations. The local-scale ROMS grid (plot every 8 grid points) varies from 600 m over the RIS and BIS to ~1 Km along the boundaries, and the regional-scale ROMS grid (plot every 4 grid points) is uniform with a resolution of 5 Km.

Figure 2-4: Comparison of depth-averaged currents in four seasons between the model (blue arrows) and observations (red arrows) in Front-Resolving Observational Network with Telemetry Project between 2000 and 2001 (Ullman and Codiga, 2004; red arrows).

Figure 2-5: Comparison of velocities across a transect along the Race (from New London, CT to Orient Point, NY) between the model (left panel) and the observation (right panel). Positive value represents current moving out of LIS, and negative value into LIS. Model-simulated salinity along the transect is shown by black contours on the right panel.

Figure 2-6: Comparison of vertical currents in spring between (left) the observations in the FRONT project (Codiga, 2004) and (right) model results. Color represents water depth with red representing currents at the top layers and blue at the deep, and arrows represent velocities at the corresponding depth.

Figure 2-7: Plan view of the surface salinity at the day of 10, 40, 70 and 100 for the case (left) Buoy and (right) RivTides.

Figure 2-8: Time averaged surface salinity for the case RivTides after 8 months' evolution and the Lagrangian trajectories of the drifters released at the mouth of LIS (marked by stars) at the first day of the 8th month and simulated with 1-hour interval.

Figure 2-9: Cross-shelf sections of (left) along-shelf and (right) across-shelf velocities along the section outside BIS, normal to the south shore of Long Island (see Figure 1) for the cases of Buoy, RivTides and RivTides_500. The white lines represent salinity, and the thick gray lines represent the isolines of 0.

Figure 2-10: Cross-shelf section of the downshelf velocities based on (color) the direct model output and (isoline) the thermal wind relation with the bottom velocity approximately set as 0. The thick gray line represents the zero isoline of the direct model output.

Figure 2-11: Momentum terms in the (left) alongshelf and (right) across-shelf directions at 15 km offshore for the cases of (black) RivTides, (blue) HeatW and (red) HeatS. The thick solid lines represent pressure gradient, the thick dashed lines represent Coriolis term, the thin solid lines represent alongshelf u-momentum advection, and the thin dashed lines represent vertical viscosity.

Figure 2-12: Seasonal variability of surface salinity for the experiment Real and the Lagrangian trajectories of the drifters released at the first day of each season.

Figure 2-13: Seasonal variability of the buoyant plume front for the case Real.

Figure 2-14: Seasonal variability of the buoyant plume front for the case NoWind.

Figure 2-15: Plane view of surface salinity after 3 days of wind forcing and the Lagrangian trajectories of the drifters released at the locations marked by stars at the first day of wind forcing and simulated with 1-hour interval for the cases driven by a). Onshore Winds (WndOn), b) Offshore Winds (WndOff), c) Upwelling Winds (WndUp) and d) Downwelling Winds (WndDown).

Figure 2-16. Same with Figure 9 but for cases WndOn, WndOff, WndUp and WndDown.

Figure 2-17: Surface salinity for the cases of (top) StratW and (bottom) StratS after 2 months of surface heating

Figure 2-18: Same with Figure 9 but for cases HeatW and HeatS.

Figure 2-19: Pressure gradients contributed by (solid lines) sea level variation and (solid lines) density gradients, i.e., sea surface elevation in the (left) alongshelf direction (right) across-shelf direction for the cases of (black) RivTides, (blue) HeatW and (Red) HeatS.

Figure 2-20: Daily-averaged (top) Surface salinity, and (bottom) surface current and vorticity (left) before, (middle) during and (right) after freshwater patches detachment. They are corresponding to the day of 186, 192 and 198, respectively.

Figure 2-21: Surface salinity (PSU) CTD surveys on spring tides during 23-25 April, 2001 (left) and during 24-26 September, 2001 by O'Donnell and Houk (2009).

Figure 2-22: Time series of Simpson number (S_i) at stations A, B and C (marked as white dots in Figure 20). The top dashed line represents the threshold between SIPS and permanently stratified stages, and the bottom represent the threshold between the well-mixed and the SIPS stages.

Figure 2-23: Time series of (top) free surface elevation, (middle) top-to-bottom salinity difference, and (bottom) vertically integrated buoyancy flux at station A close to the Connecticut River plume front. The blue lines represent hourly averaged results, and red lines are low pass filtered results with cutoff frequency of 60 hours.

Figure 2-24: As in Figure 23 but for station B.

Figure 2-25: As in Figure 23 but for station C.

Figure 2-26: Freshwater flux across the section between Montauk Point and Block Island (section 3 in Figure 1) from the day 181 to the day 240 for the case RivTides. The three gray lines represent the days of 186, 192 and 198, which represent a time before, during and after the eddy detachment, respectively.

Figure 2-27: Momentum terms (top) in Xi (i.e., along-shelf) direction at the surface of section E and (bottom) in Eta (i.e., across-shelf) direction at the surface of section X during the detachment. The locations of the sections are marked in Figure 20.