Tidal-, Wind-, and Buoyancy-Driven Dynamics in the Barataria Estuary and Its Impact on Estuarine-Shelf Exchange Processes

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TIDAL-, WIND-, AND BUOYANCY-DRIVEN DYNAMICS IN THE BARATARIA ESTUARY AND ITS IMPACT ON ESTUARINE-SHELF EXCHANGE PROCESSES

A Dissertation

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Doctor of Philosophy

in

The Department of Oceanography and Coastal Sciences

by

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SYMBOLS

\( R_0 \)  
Rossby number, also denotes internal Rossby radius

\( E_k \)  
Ekman number

\( K \)  
Kelvin number

\( A_z \)  
Vertical eddy viscosity

\( H \)  
Water depth

\( B \)  
Width of the channel

\( f \)  
Coriolis parameter

\( U \)  
Estuarine circulation velocity

\(|u|\)  
Absolute value of lateral velocity amplitude

\( \sigma \)  
Semidiurnal tidal frequency

\( L \)  
Plume length in alongshore extend

\( g' \)  
Reduced gravity acceleration

\( R_i \)  
Gradient Richardson number

\( F_r \)  
Froude number
ABSTRACT

A three-dimensional, high-resolution, Finite-Volume Coastal Ocean Model (FVCOM) was used to study the dynamics of Barataria Estuary located in the Southeastern Louisiana. Three numerical experiments with different discharge scenarios, including the actual discharge (average ~ 160 m$^3$ s$^{-1}$) from the Davis Pond Diversion (DPD) over three months from April to June 2010, no discharge (NO), and the proposed Mid-Barataria Diversion (MBD) with a constant discharge of 850 m$^3$ s$^{-1}$, were conducted to investigate the impacts of river diversions on salinity gradients and residence times in the estuary. The three-month average salinity indicated that surface salinity had less variation in the DPD scenario compared with that in the NO scenario, while bottom salinity differences between the DPD and the NO scenarios were as high as ~ 4. On the other hand, the maximum average salinity for both surface and bottom in the MBD scenario exhibited a reduction of ~ 12 compared with that in the NO scenario, with a larger area at the bottom than at the surface. Both the DPD and the MBD had a great impact on the residence time of Barataria Bay, where the average residence time was reduced from 15 days in the NO scenario, to 6 (4) days in the DPD (MBD) scenario, when passive particles were released at flood slack.

Barataria Pass is one of the four tidal inlets connecting the Barataria Bay with the coastal ocean and has the greatest estuarine-shelf exchange. The lateral circulation in the inlet showed a pair of counter-rotating circulation developing during flood tide, while unidirectional flow occurred during ebb tide. Analysis of 3-D momentum equations revealed that nonlinear advection is the dominant force generating lateral circulation in this narrow inlet. Model results showed that ebb tides transported freshwater seaward through this inlet to form radially spreading estuarine plumes over the adjacent continental shelf. Wind-driven coastal circulation determined the subtidal variations of the plume when the upstream freshwater discharge rate was
almost constant. Particles released near the seaward side of the inlet at ebb slack could be transported into the bay, however, most of them were expelled out during the next ebb tides and floated over the convergence zone of the plume front.
CHAPTER 1. INTRODUCTION

Since the last century, coastal Louisiana has experienced serious land loss, as a result of both natural and man-made factors. Natural factors can be attributed to subsidence, erosion and sea level rise (Boesch et al., 1994; Day et al., 2000), as well as hurricane induced storm surge, which can easily convert wetland to open water (Day et al., 2007). On the other hand, levee and dam construction, along with dredging of oil and gas canals are man-made factors. All these factors have accelerated land loss, with the rate reaching a maximum of 102 km² y⁻¹ in the 1970s (Barras et al., 2003). Even though the rate has slowed down, land loss is still one of the most significant problems for coastal Louisiana.

Barataria Basin is located in the west of the lower Mississippi River. It is bounded to the east by the Mississippi River, to the west by Bayou Lafourche, and in the south is connected to the coastal ocean through multiple tidal inlets. Artificial levees were constructed along the western and eastern side of the basin (Conner and Day, 1987). As a result, sediment supply for Barataria Basin is limited. Freshwater diversions from the Mississippi River are considered to be the most effective and inexpensive restoration methods. River diversions can divert sediment-laden water through distributary channels into adjacent wetland basins. The Atchafalaya River can be regarded as a successful example of using river diversions to build lands (Roberts, 1998; Rosen and Xu, 2013). Also, river diversions have been used in a number of deltas for different purposes (Allison et al., 2014).

The existing diversion in the Barataria Estuary is the Davis Pond freshwater diversion, which is located on the west bank of the Mississippi River. The diversion structure consists of four iron-gated 4.3-m by 4.3-m box culverts built into the levee. There is also a 163 m long by 26 m wide inflow channel. An outflow channel connects the structure to the ponding area and
diverts freshwater into the estuary. The Davis Pond diversion was opened in 2002 and pumps an average of 140 m$^3$ s$^{-1}$ of water into the basin. The maximum capacity is ~ 300 m$^3$ s$^{-1}$. The primary purpose of this freshwater diversion is to imitate spring flooding to provide controlled flow of freshwater, sediments and nutrients from the Mississippi River to the Barataria Estuary.

The Mid-Barataria Sediment Diversion is one of the sediment diversions proposed in the 2012 and 2017 Coastal Master Plans (CPRA, 2012, 2017), and is located along the west bank of the Mississippi River near Myrtle Grove. It is one of the largest coastal and ecosystem restoration projects, aiming to reconnect the river to the impacted area and divert sediments to build new lands. This project is still under the permit review process due to its unprecedented size and scope. A detailed and broad environmental impact statement must be developed prior to its construction.

A salinity gradient is an important factor that can influence the spatial pattern of biota and biogeochemical processes. For example, using long-term measurements of chemical, biological, and physical properties along the salinity gradient of San Francisco Bay, Cloern et al. (2017) found that the geographic location of motile estuarine species (plankton, crustaceans, fish) track the salinity gradient. The classic salinity zonation scheme is the Venice System, which divides salinity zone into six categories, 0-0.5, 0.5-5, 5-18, 18-30, 30-40, >40 (Anonymous 1958). Greenwood (2007) assessed the nekton community change along estuarine salinity gradients in southwest Florida estuaries and found that the nekton community structure changes rapidly at each salinity end zone, with relatively slow change in between zones. Similar research was conducted in lower St. Johns River estuary (Guenther and MacDonald, 2012). Christensen et al. (1997) developed an index to assess the sensitivity of Gulf of Mexico species to changes in estuarine salinity regimes and found a significant difference exists between adult and juvenile
life stage sensitivity, with juveniles exhibiting a lower sensitivity to salinity changes than adults. They also documented a great disparity in species-specific sensitivities among Gulf estuaries.

The salinity gradient of estuaries also plays an important role in determining spatial patterns of physical properties. Stratification is a distinct feature of estuaries caused by salinity differences between the surface and bottom of the estuary, which can inhibit the vertical motion. Estuarine residence time is “the time it takes for any waterparcel of the sample to leave the lagoon through its outlet to the sea” (Dronkers and Zimmerman, 1982). This parameter represents the timescales of physical transport processes and can be used to compare with the timescales of biogeochemical processes to assess nutrient exports and imports (Dettmann, 2001), to estimate algal biomass (Vollenweider, 1976) and dissolved organic carbon concentration (Christensen et al., 1996), and to estimate primary production (Jassby et al., 1990). Due to the importance of these two factors, an investigation of residence time variations caused by river diversions is imperative and urgent. In chapter 2, a three-dimensional, high-resolution hydrodynamic model is used to estimate the impact of Mississippi River diversions on the salinity gradient and residence time.

Lateral circulation was observed in a well-mixed estuary in North Wales by Nunes and Simpson (1985). They found an axial surface front during flood tides. They then constructed a diagnostic model, which predicted a pair of counter-rotating circulation cells in the cross-channel section, showing converge at the surface and divergence at the bottom. Based on the model, they postulated that this lateral circulation was generated by differential advection of along-channel density gradients, resulting in a larger density in the channel than that in the shoals. During the last two decades, lateral circulation has attracted the attention of the scientific community and inspired a number of investigations aimed at identifying its driving mechanisms, based on
numerical modeling (Chen and Sanford, 2009; Lerczak and Geyer, 2004) and observations (Chant and Wilson, 1997; Lacy and Monismith, 2001; Li, 2002; Nidzieko et al., 2009; Scully et al., 2009). Except for differential advection, other driving forces can generate lateral circulation, including boundary mixing on a sloping bottom (Chen and Sanford, 2009), Coriolis deflection (Huijts et al., 2009; Scully et al., 2009), interactions between barotropic tides and cross-channel variations in bathymetry (Li, 2002; Li and Valle-Levinson, 1999; Valle-Levinson et al., 2000), and flow curvature (Lacy and Monismith, 2001; Nidzieko et al., 2009). Several studies have found that lateral circulation plays an important role in estuarine circulation. Modeling in idealized, straight, stratified estuaries, Lerczak and Geyer (2004) found that lateral advection acts as a driving term for the estuarine exchange flow and can be larger than the along-channel pressure gradient in the tidally averaged, along-channel momentum equation. A similar conclusion was also found by Scully et al. (2009). Their model shows that the nonlinear advective acceleration terms, driven by secondary lateral circulation, contribute to the subtidal along-channel momentum balance at a leading order. Cheng and Valle-Levinson (2009) examined the influence of non-linear lateral advection on estuarine exchange flow using a scaling analysis. They found that the relative importance of lateral advection and the earth’s rotation on estuarine dynamics can be evaluated in terms of the nondimensional Rossby and Ekman numbers ($R_0$ and $E_k$). The classic estuarine dynamics occur at intermediate $R_0$ and large $E_k$, with vertically sheared exchange flows that outflow at the surface and inflow at the bottom. To date, lateral circulation has been ignored in the studies of Louisiana estuaries. However, when it exists, it may have a great impact on estuarine-shelf exchange. In chapter 3, lateral circulation is examined in the tidal inlet, Barataria Pass, which conveys the greatest exchange between bay and coastal ocean. The driving mechanisms are analyzed by 3-D momentum equations.
Estuary and the inner shelf are strongly coupled through the exchange of different water masses across the mouth of estuaries. The response to forcing in either the estuary or the shelf differs from systems where the two regimes are isolated from each other. The distribution of dissolved and particular matter within an estuary and the coastal ocean thus depends upon the circulation, the mixing and the dynamics of both regimes. Tidal inlets are conduits for water exchange between estuaries and inner shelf, which are relatively short, narrow channels. Inlet circulation is controlled by tides, bay geometry, inlet geometry, bottom topography, and nontidal forcing, such as river discharge and wind. Figure 1.1 shows circulation and morphological characteristics at symmetric (i.e. islands’ width are equal) inlets. Flood currents are similar to potential flow (Kapolnai et al., 1996) and drive water transverse of the inlet and into the bay. When the velocity is reduced, materials are deposited, generating two shoals on both side of the inlet channel on the bay side (Hayes, 1975). Outflows (ebb flows) exit the inlet through the channel, forming a main ebb channel (Fig. 1.1). The flow may be featured as gravity currents or a jet (Kapolnai et al., 1996). Tidal and buoyancy forces often control the circulation in an estuary-inlet-shelf system on intratidal scales, while winds can drive circulation on a synoptic scale. Small buoyancy fronts (Garvine, 1984; Wiseman and Garvine, 1995) are also found along the estuarine mouth, separating seaward flowing estuarine freshwater from saline shelf water. The persistent nature of plume fronts makes them ideal for examining their role in floating particles or fish larvae flux. Larval fish can cross the river plume front, while mixing and stirring can account for shoreward transport of larval fishes (Govoni et al., 1989). In chapter 4, tidal- and wind-driven circulations near the inlet and mixing in the inlet and estuarine plumes are discussed.

Summary and general conclusions are presented in chapter 5.
Figure 1.1. Typical inlet flow patterns and morphological characteristics (Hayes, 1975).
CHAPTER 2. IMPACT OF RIVER DIVERSIONS ON SALINITY GRADIENT AND RESIDENCE TIME IN THE BARATARIA ESTUARY

2.1. Introduction

Over the past two centuries, artificial levees were constructed along the Mississippi River to prevent flooding. Currently, levees extend about 3620 km, which protect populated and agricultural areas from flooding and maintain navigational channels (Dean et al., 2006). On the other hand, they also restrict the delivery of nutrient-rich river water and sediment to coastal wetlands. Levee and canal construction, combined with sea-level rise, subsidence, and marine processes (e.g., waves and currents) have caused a widespread loss of coastal wetlands. Controlled river diversions that create pulses of sediment-laden fresh water are increasingly being used for coastal wetland restoration. River diversions were first proposed in 1984 (Chew, 1984) to combat saltwater intrusion. Later, river diversions were proposed as a tool for marsh and barrier island restoration. Wax Lake Delta is a successful example of using river diversions to build lands (Roberts, 1998; Rosen and Xu, 2013).

The existing controlled river diversion in Barataria Estuary is Davis Pond Diversion (Fig. 2.1). It was completed in 2002 and is located at river mile 118, above Head Pass on the west bank of the Mississippi River. The diversion structure consists of four, 4.3-m by 4.3-m gated box culverts. The capacity is ~ 300 m$^3$ s$^{-1}$ (Das et al., 2012). Freshwater discharge began in July of 2002 and became fully operational in 2009. The primary motivation of Davis Pond Diversion was to enhance marsh vegetation growth, combat marsh loss and increase estuarine productivity by introducing freshwater and associated nutrients from the Mississippi River into the basin (Plitsch, 2014). In addition, river diversions are considered to be the most efficient way to build lands (Boesch et al., 1994). To this end, the Coast 2050 report proposed constructing a delta-
building diversion in Myrtle Grove/Naomi area, which is expected to build land and prevent land loss in the central basin by 2050 (COAST2050, 1998). The Mid-Barataria Diversion has been studied with observational data and numerical modeling (Allison et al., 2014; Allison et al., 2017; Meselhe et al., 2012) and is the first sediment diversion approved to be built in the Barataria Basin. This diversion will convey a maximum of ~2100 m$^3$ s$^{-1}$ of freshwater. However, one of the controversies that arise with diversions is their impact on aquatic biota. The previous investigations mostly focus on sediment dynamics. Little is known about how freshwater will affect hydrodynamics, such as the salinity gradient and residence time. Diverting massive freshwater into the system has a potential to dramatically influence the salinity gradient, as well as turbidity and water quality. Based on scenarios with actual freshwater discharges in different years and scenarios with several new diversions, Das et al. (2012) found that river diversions can strongly affect salinities in the middle section of the Barataria Estuary. The maximum discharge in their scenarios is ~300 m$^3$ s$^{-1}$. The proposed Mid-Barataria Diversion will convey orders of magnitude larger freshwater to the Barataria Basin than the existing Davis Pond Diversion and is located in the central of the basin. Thus, it may cause significant displacement of salinity stress to commercially and recreationally important fish and shellfish species due to the wide-spread and prolonged freshening of habitat. Diversions will also increase nutrient loading, which may increase phytoplankton production and promote the development of noxious or toxic phytoplankton blooms in the estuarine environment. Estuarine water residence time is often used as an indicator for phytoplankton biomass accumulation, as blooms are less likely to develop in systems with a short residence time.

The objective of this study is to provide a quantitative assessment of the impacts of existing and proposed river diversions on the salinity gradient and residence time in the Barataria
Estuary based on different discharge scenarios. A three-dimensional, high-resolution hydrodynamic model was built by our team and first used in this area.

Figure 2.1. Map of the Barataria Estuary (black line bordered) showing the Davis Pond Diversion, proposed Mid-Barataria Sediment Diversion, and water bodies of interest in this study.

2.2. Study area

The Barataria Estuary is located southwest of New Orleans, Louisiana (Fig. 2.1). It is roughly triangular in shape, about 110 km north-south and 50 km wide at its mouth, where it is separated from the Gulf of Mexico by chains of barrier islands. Morphological features are characterized by natural and artificial levees, bays, lakes, bayous, barrier islands, and swamp and marsh wetlands. Due to subsidence and erosion, bays, lakes, and bayous have enlarged to form an extensive network of interconnecting waterbodies which allow for transport of water, materials and migrating organisms throughout the estuary (Conner and Day, 1987). Barrier islands protect the estuary from waves and currents that would otherwise worsen the already
rapid erosion of wetlands. The average depth of the estuary is about 2 m (Das et al., 2012), excluding the navigation channel. The deepest area is ~ 50 m and is located near the Barataria Pass. The main sources of freshwater are from rainfall, stream runoff, freshwater diversions and siphons, and the Gulf Intracoastal Waterway. The tide is predominantly diurnal and small, with a tropic-equatorial cycle, where tidal range is approximately 0.35 m during tropic tides near the mouth (Das et al., 2012) and is attenuated gradually toward the upper estuary. The Barataria Estuary is connected to the coastal oceans by four main tidal inlets, including Caminada Pass, Barataria Pass, Pass Abel, and Quatre Bayou Pass.

2.3. Methods

2.3.1. Hydrodynamic model configuration

The Finite Volume Coastal Ocean Model (FVCOM, version 2.6) was used in this study to simulate the hydrodynamics of the Barataria Estuary. FVCOM employs a horizontal triangular grid, which can resolve complex coastlines and flow geometries. A wet/dry point treatment method was incorporated into FVCOM. If vertical water column thickness was less than 5 cm, the cell was designated as a dry cell and its velocity was set to zero. When vertical water column thickness exceeded 5 cm, the water level and velocity were computed from discrete equations.

The computational domain consisted of 146,266 triangle nodes and 283,721 triangular cells, extending longitudinally from east of Mobile Bay, Alabama to west of Galveston Bay, Texas, and offshelf to about 27°N (Fig. 2.2), with inclusion of the intertidal zone inside the Barataria Bay. The horizontal grid resolution varied from ~ 10 m in the upper estuary to ~ 8 km near the open boundary. The finest grid cells were small enough to represent the narrow channels with three continuous grids to ensure the simulation of mass transport along the channels. The
model included 18 sigma layers evenly distributed over the water column. Computational time steps were 0.3 s and 3 s for the external and internal modes, respectively. Model results were saved every hour for further analysis.

Bathymetry was obtained from various sources. Using an inverse distance weighted interpolation method, a 5 m by 5 m resolution digital elevation model (DEM) constructed from Light Detection and Ranging (LIDAR) measurement was interpolated into wetland region of the domain. The water depth in channels, bayous, and lakes in the estuary was interpolated from Coastal Louisiana Ecosystem Assessment and Restoration Report (CLEAR), US Army Corps Survey, and NOAA nautical charts. The water depth values in the Barataria Pass were obtained from vessel-based surveys (Li et al., 2011). Shelf and open ocean water depth were interpolated from a coarse resolution ADCIRC model bathymetry. The interpolated model depth is shown in Fig. 2.2.

2.3.2. Lagrangian particle tracking

The FVCOM offline Lagrangian particle tracking module solves a non-linear system of ordinary differential equations (Chen et al., 2011). The rate of change of particle position is described by

\[
\frac{d\vec{x}}{dt} = \vec{v}(\vec{x}(t), t)
\]

where, \(\vec{x}\) is the position of given particle, and \(\vec{v}(\vec{x}, t)\) is the velocity field taken from the FVCOM hydrodynamic model, which is bi-linearly interpolated in space under Cartesian coordinates and linearly interpolated in time. This Lagrangian solver is solved by a fourth-order explicit Runge-Kutta time-integrator. In this study, only the horizontal advection tracking program was included.
Figure 2.2. Bathymetry used in the numerical model. (a) Local domain of Barataria Estuary, T1-T5 are transects used to calculate water flux, (b) entire model domain.
2.3.3. Initial and boundary conditions

In this study, the model was initialized on 00:00 April 1, 2010 (GMT), and ran until June 30, 2010. The fluid is initially at rest, and the model spins up with a ramping period of 10 days. The model requires specification of input parameters for initial conditions (sea level elevation, velocity, and salinity), river boundary condition (river discharge), surface boundary condition (wind stress), and open boundary conditions (sea level elevation).

The initial sea level elevation and velocity were specified as zero throughout the entire computational domain. Initial salinity was interpolated from two data sources. Salinities in the Barataria Estuary were interpolated from observations at the USGS stations, while salinities over the continental shelf were interpolated from HYCOM Gulf of Mexico 1/25° Reanalysis product (https://hycom.org/data/goml0pt04/expt-20pt1). Vertical salinity in the Barataria Estuary was uniformly distributed. Experiments show that this technique gives more accurate salinity simulation result than with linear interpolation from estuarine head to the mouth.

Observed 15-min freshwater discharge at four locations (Mississippi River at Belle Chasse, Atchafalaya River at Morgan City, Wax Lake, and Davis Pond diversion) were injected into the computational domain with flux boundary conditions of zero salinity and specified volume and momentum. Fig. 2.3 shows the time series of discharges for the Mississippi River and Davis Pond Diversion in 2010. The Mississippi River discharge was \( \sim 35,000 \, \text{m}^3 \, \text{s}^{-1} \) during flooding period. During this simulation period, discharge varied from 12,000 to 26,000 \( \text{m}^3 \, \text{s}^{-1} \). The peak period for Davis Pond Diversion lasted from May to August in 2010, with discharge up to 300 \( \text{m}^3 \, \text{s}^{-1} \), which was close to its capacity.
Figure 2.3. Time series of 15-min interval discharge for the Mississippi River at Belle Chasse (black line) and the Davis Pond Diversion (red line) in 2010.

Three-hourly, 10-m wind data was obtained from NOAA National Centers for Environmental Prediction (NECP) North American Regional Reanalysis (NARR) products and interpolated onto the entire computational domain using a distance-square-weighted algorithm with two points in each quadrant. The effects of precipitation and evaporation were not considered in this study.

Sea level elevations at the open boundary nodes were estimated by extrapolation method. Specifically, the 6-min sea level elevations at four locations were downloaded from NOAA tides and currents website (https://tidesandcurrents.noaa.gov/). Time series of Dauphin Island, Southwest Pass, Freeport, and Galveston Pier 21 were directly used to prescribe sea level elevations at the easternmost node, the southeastern node, the southwestern node, and the westernmost node, respectively. Huang et al. (2011) recently utilized this method in the adjacent Breton Sound Estuary and it performed well, which gives us confidence to use it in this study.
2.3.4. Model scenarios

Three different diversion scenarios were chosen to analyze the impact of river diversions on the salinity gradient and residence time. The initial and open boundary forcing functions were identical for all model scenarios. The "DPD" experiment used the actual 2010 Davis Pond diversion. The "NO" experiment turned off the Davis Pond diversion. The "MBD" experiment turned off the Davis Pond diversion and opened the proposed Mid-Barataria sediment diversion, in which the discharge was set to ~ 850 m$^3$ s$^{-1}$ over the entire simulation period. This discharge value was used to satisfy the stability of the model as a discharge greater than ~ 850 m$^3$ s$^{-1}$ resulted in an unstable model. In the MBD scenario, the Mississippi River discharge was reduced by ~ 850 m$^3$ s$^{-1}$.

2.3.5. Observational data

Time series of sea level elevation data were obtained from NOAA tides and currents products, including Gulfport, Bay Waveland, Southwest Pass, Grand Isle, Calcasieu Pass, and Sabine Pass, as well as from the National Water Information System of USGS, including Barataria Waterway, North of Grand Isle, Hackberry Bay, Bay Dosgris, Little Lake at Cutoff and Cataouatche Lake. Salinity measurements were also obtained from USGS stations. For this study, there was no observed velocity data for model validation.

2.4. Model validations

Correlation coefficients (CC), root-mean-square-errors (RMSE), and indexes of agreements (IOA) were used for model validation, which are given as (Willmott, 1981):
where $x$ and $y$ are observed and simulated values, $\bar{x}$ and $\bar{y}$ are the mean value of $x$ and $y$, respectively, and $n$ is the size of data. The IOA values range from 0 to 1, with 1 representing a perfect agreement between model results and observations. The mean values and standard deviations are also given for observation and simulation, respectively. Quantitative comparisons between observed and simulated sea level elevations are given in Tab. 2.1.

Simulated sea level elevations demonstrated strong agreement with observations, both for NOAA stations (Fig. 2.4) and USGS stations (Fig. 2.5). The correlation coefficients for all USGS stations and most NOAA stations were above 0.9 (Tab. 4.1). The two stations with low correlation coefficients were located on the western side of the model domain. These low correlation coefficients may be due to the location difference between model and observation, as can be seen in Fig. 2.4f and Fig. 2.4g, the flood tides are retarded in the observations. The indexes of agreement for most stations exceeded 0.9. The two stations with relatively low IOAs were Southwest Pass and Grand Isle, whose values were below 0.8. However, their correlation coefficients were very high, with 0.95 and 0.93, respectively. The low IOAs may be caused by vertical datum difference between observation and model (Fig. 2.4d and Fig. 2.4e).

Statistical assessments of surface salinities are given in Tab 2.2. Time series of modeled and observed surface salinities are shown in Fig. 2.6. The 10-day spin-up time seems to be
reasonable in this simulation, as simulated salinities are very close to the observations after 10-day adjustment (Fig. 2.6). The model captures the variations of surface salinity. The sudden rises of salinity on April 24 and May 2 are also reproduced. This increase may be related to salt intrusion induced by southeastern wind (Fig. 2.5a). Generally, the model overestimates the salinity, since simulated mean values are greater than mean observations (Tab. 2.2). However, during two wind events, the simulated surface salinities are lower than observed ones (Figs. 2.6d, 2.6e, 2.6f). The correlation coefficients for most stations are above 0.7 except Barataria Waterway. Salinity differences between model and observation in June (Day 151 -181) could be caused by several reasons. The mean discharge at USGS GIWW West of Bayou Lafourche at Larose is ~ 50 m$^3$ s$^{-1}$, which is not included in our model. Also, several small river diversions in the lower estuary are not included in the model. No numerical model can be broken away from model error and uncertainties. Similarly, observational data may also contain intrinsic errors, spatial variability, and uncertainty.

Table 2.1. Statistical assessment of model performance for sea level elevations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Observed Mean (m)</th>
<th>Observed Std. (m)</th>
<th>Simulated Mean (m)</th>
<th>Simulated Std. (m)</th>
<th>CC</th>
<th>RMSE (m)</th>
<th>IOA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulfport</td>
<td>0.11</td>
<td>0.23</td>
<td>0.16</td>
<td>0.21</td>
<td>0.93</td>
<td>0.09</td>
<td>0.95</td>
</tr>
<tr>
<td>Bay Waveland</td>
<td>0.14</td>
<td>0.23</td>
<td>0.17</td>
<td>0.19</td>
<td>0.91</td>
<td>0.10</td>
<td>0.94</td>
</tr>
<tr>
<td>Southwest Pass</td>
<td>0.15</td>
<td>0.16</td>
<td>0.34</td>
<td>0.18</td>
<td>0.95</td>
<td>0.20</td>
<td>0.75</td>
</tr>
<tr>
<td>Grand Isle</td>
<td>0.06</td>
<td>0.16</td>
<td>0.22</td>
<td>0.14</td>
<td>0.93</td>
<td>0.16</td>
<td>0.78</td>
</tr>
<tr>
<td>Calcasieu Pass</td>
<td>0.10</td>
<td>0.22</td>
<td>0.23</td>
<td>0.32</td>
<td>0.84</td>
<td>0.22</td>
<td>0.83</td>
</tr>
<tr>
<td>Sabine Pass</td>
<td>0.10</td>
<td>0.19</td>
<td>0.22</td>
<td>0.32</td>
<td>0.85</td>
<td>0.22</td>
<td>0.82</td>
</tr>
<tr>
<td>Barataria Waterway</td>
<td>0.33</td>
<td>0.15</td>
<td>0.26</td>
<td>0.12</td>
<td>0.92</td>
<td>0.09</td>
<td>0.89</td>
</tr>
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<td>North of Grand Isle</td>
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<td>0.13</td>
<td>0.90</td>
<td>0.09</td>
<td>0.92</td>
</tr>
<tr>
<td>Hackberry Bay</td>
<td>0.19</td>
<td>0.16</td>
<td>0.25</td>
<td>0.13</td>
<td>0.92</td>
<td>0.08</td>
<td>0.92</td>
</tr>
<tr>
<td>Bay Dosgris</td>
<td>0.26</td>
<td>0.12</td>
<td>0.26</td>
<td>0.12</td>
<td>0.92</td>
<td>0.05</td>
<td>0.96</td>
</tr>
<tr>
<td>Little Lake at Cutoff</td>
<td>0.31</td>
<td>0.13</td>
<td>0.29</td>
<td>0.11</td>
<td>0.92</td>
<td>0.05</td>
<td>0.95</td>
</tr>
<tr>
<td>Cataouatche Lake</td>
<td>0.35</td>
<td>0.15</td>
<td>0.34</td>
<td>0.12</td>
<td>0.96</td>
<td>0.05</td>
<td>0.96</td>
</tr>
</tbody>
</table>
Table 2.2. Statistical assessment of model performance for surface salinity.

<table>
<thead>
<tr>
<th>Station</th>
<th>Observed</th>
<th></th>
<th>Simulated</th>
<th></th>
<th>CC</th>
<th>RMSE</th>
<th>IOA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>Std.dev.</td>
<td>Mean</td>
<td>Std.dev.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>North of Grand Isle</td>
<td>5.6</td>
<td>4.2</td>
<td>9.5</td>
<td>2.7</td>
<td>0.73</td>
<td>4.8</td>
<td>0.68</td>
</tr>
<tr>
<td>Hackberry Bay</td>
<td>5.0</td>
<td>4.1</td>
<td>7.7</td>
<td>1.9</td>
<td>0.73</td>
<td>4.0</td>
<td>0.68</td>
</tr>
<tr>
<td>Barataria Waterway</td>
<td>4.9</td>
<td>4.1</td>
<td>7.3</td>
<td>1.3</td>
<td>0.58</td>
<td>4.2</td>
<td>0.58</td>
</tr>
<tr>
<td>Bay Dosgris</td>
<td>4.7</td>
<td>4.7</td>
<td>7.5</td>
<td>2.5</td>
<td>0.77</td>
<td>4.2</td>
<td>0.73</td>
</tr>
<tr>
<td>Little Lake near Cutoff</td>
<td>2.5</td>
<td>3.1</td>
<td>2.7</td>
<td>0.7</td>
<td>0.70</td>
<td>2.6</td>
<td>0.49</td>
</tr>
</tbody>
</table>

Figure 2.4. (a) Time series of wind vector at Grand Isle and (b-g) comparison of 10-min observed (solid) and hourly simulated (dashed) sea level elevations at six coastal NOAA stations.
Figure 2.5. (a) Time series of wind vector at Grand Isle and (b-g) comparison of hourly observed (solid) and simulated (dashed) sea level elevations at six USGS stations in Barataria Estuary. Shaded areas indicate missing values.
Figure 2.6. Comparison of hourly observed (red dots) and modeled (black line) salinity values at six USGS stations in Barataria Estuary. Shaded area denotes 10-day spin-up period.
2.5. Discussion

2.5.1. Impact of diversions on residual currents

Shown in Fig. 2.7 are three-month mean surface currents for the upper estuary (Fig. 2.7a-c), middle estuary (Fig. 2.7d-f), and lower estuary (Fig. 2.7g-i), respectively. The bottom residual currents are shown in Fig. 2.8. Compared to the no diversion scenario, the DPD had a greater impact on wetland inundation within a radius of about 10 km from the diversion structure at both the surface and bottom (Fig. 2.7a, Fig. 2.8a). Surface residual currents in the channel along the freshwater pathway were also be affected by the DPD. As shown in Fig. 2.7d, velocities in the Little Lake near Bay Dosgris increased. The DPD had little impact on residual currents near the mouth of the estuary, while the MBD, affected most areas on the eastern side of the Barataria Waterway, with the largest currents converged along the channel (Fig. 2.7f). The greatest inundation occurred within a radius of about 25 km from the diversion structure at both the surface and bottom (Fig. 2.7f and Fig. 2.8f). The surface residual currents in the Little Lake (Fig. 2.7f) and estuarine mouth (Fig. 2.7i) also greatly increased. The outflows moved further seaward. The mean velocity in the Barataria Waterway greatly increased in the MBD scenario, with the magnitude increase reaching up to 0.4 m s\(^{-1}\) at the surface and 0.1 m s\(^{-1}\) at the bottom (Fig. 2.7f, Fig. 2.8f). This indicates that freshwater from the MBD can go through upstream tributaries to reach Barataria Waterway.
Figure 2.7. Simulated three-month average surface currents for (a-c) the upper estuary, (d-f) middle estuary, and (g-i) lower estuary. Current vectors were selected in a search radius of ~ 500 m and with velocities between 0.01 m/s and 0.3 m/s. The first column is the DPD scenario, the second column is the NO scenario, and the third column is the MBD scenario.
Figure 2.8. Same as Fig. 2.7, but for bottom currents.
2.5.2. Impact of diversions on estuarine salinities

Time series of observed and simulated surface salinities for different scenarios are shown in Fig. 2.9. To examine how diversions affect salinities near the estuarine mouth, simulated salinities near Grand Isle are also shown in Fig. 2.9a. Results indicate that both the DPD and MBD have little impact on salinity at Cataouatche Lake (Fig. 2.9g) and Little Lake near Cutoff (Fig. 2.9f). This is because parts of the upper estuary are fresh most of the time, where the mean salinities are below 5. The 2-D model results also had similar findings (Das et al., 2012). Based on a 2-D coupled hydrology-hydrodynamic box model simulation in 2002, Das et al. (2012) found that the salinity differences can be as high as 10 between different discharge scenarios within the Little Lake area. Our model shows that the salinity differences between the DPD and the NO are less than 5 in the Little Lake near Dosgris, during fair conditions in 2010 (Fig. 2.9e).

For North of Grand Isle (Fig. 2.9b), Hackberry Bay (Fig. 2.9c), and Barataria Waterway (Fig. 2.9d), both DPD and MBD decreased salinities at these stations, however, the salinity differences between the DPD and the MBD were relatively small in fair conditions. These results indicate that even though the magnitude of discharge is much greater than that of the DPD, the MBD had little impact on surface salinities over most estuarine areas. It is important to note that during two saltwater intrusion events, the DPD had little impact on surface salinities, while the MBD produced a great reduction in surface salinities. For example, salinity differences between the DPD and the MBD at Hackberry Bay was as great as 10 over these two events (Fig. 2.9c). The most affected area was near Grand Isle, where tidal exchange accounted for approximately 85% of the flow variability (Snedden, 2006). Surface salinities in the MBD scenario had the largest range, with low salinity as low as ~ 8 and as high as ~ 20, with large reductions occurring during ebb tides.
Figure 2.9. Observed (red dots) and simulated surface salinity for different diversion scenarios for April – June, 2010. DPD (black line) denotes the scenario with a realistic Davis Pond diversion. NO (blue line) denotes the scenario without diversion. MBD (green line) denotes the scenario with 850 m$^3$ s$^{-1}$ discharge from the Mid-Barataria Diversion.
With the opening of diversions, freshwater would increase the outflows and greatly decrease salinities during ebb tides. Fig. 2.10 shows distribution of simulated surface salinities at 08:00 May 17, 2010 (ebb tide) under different diversion scenarios (Fig. 2.10a, 2.10b, 2.10c) and their differences (Fig. 2.10d, 2.10e). As expected, the lower estuary had the greatest difference. Adjacent areas within a radius of about 10 km from the DPD structure were flooded (Fig. 2.10a). The simulated salinity difference between the NO and the DPD scenarios were as great as 6 (Fig. 2.10d), mainly near Barataria Pass, which is the main exchange inlet between Barataria Bay and open oceans. Salinities in the Little Lake and Hackberry Bay also experienced variability. With 850 m$^3$ s$^{-1}$ discharge, the MBD inundated areas within a radius of about 25 km from the structure (Fig. 2.10c). It reduced instantaneous surface salinities across a large portion of the estuary, from Lake Salvador to middle estuary to coastal ocean, with salinity differences varying from ~ 2 to ~16 (Fig. 2.10e). Outflows transported freshwater westward, which also resulted in surface salinity reduction (~ 2) in the Terrebonne Bay (Fig. 2.10e).

The three-month average salinity difference between the NO and DPD scenarios and between the NO and MBD scenarios at both the surface and bottom are shown in Fig. 2.11. The Davis Pond Diversion had little impact on surface mean salinities (Fig. 2.11a), but reduced bottom mean salinities in the middle and lower estuary, with salinity differences up to ~ 3 (Fig. 2.11c). These relatively sensitive areas are consistent with previous findings that the mid-estuary may be the most influenced by potential diversion projects (Inoue et al., 2008; Wang et al., 2016) because of their steep salinity gradients. Like the DPD, the MBD had minimal impact on upper portions of the estuary because these areas are predominantly fresh, with low salinity values. However, the MBD had a large impact on mean salinities in areas from the mid-estuary to coastal ocean. The surface mean salinities in the Barataria Bay and Caminada Bay were reduced
by ~ 4 on average, but a small portion decreased by as much as ~ 12 (Fig. 2.11b). This is consistent with recent modeling results (White et al., 2018) showing that the MBD has much larger impacts on periodic freshening near Caminada Bay. In addition, a large portion of bottom areas were impacted by the MBD. The bottom mean salinities in the Barataria Bay were reduced by ~ 12. The MBD also caused bottom mean salinity reduction over coastal ocean and Terrebonne Bay, with salinity difference ~ 2.

Figure 2.10. Simulated surface salinity at 08:00 May 17, 2010 (GMT) for (a) the DPD scenario, (b) NO scenario, (c) MBD scenario and salinity difference between (d) NO and DPD scenarios and (e) between NO and MBD scenarios.
Figure 2.11. Three-month average salinity differences between (a) the NO and DPD scenarios at the surface, (b) the NO and MBD scenarios at the surface, (c) the NO and DPD scenarios at the bottom, and (d) the NO and MBD scenarios at the bottom.
2.5.3. Impact of diversions on residence time

Following Dronkers and Zimmerman (1982), the residence time in this study is defined as the average time that any water parcel (represented by multiple Lagrangian particles) takes to leave the estuary through its outlet to the sea. Residence time is measured from an arbitrary start location within the waterbody. It is affected by initial position (Fig. 2.12), tidal and wind circulations, diversion discharge and releasing time. In a typical shallow estuary with extremely complex geomorphology, particles are likely to be trapped or flushed onto the land at certain times. Recognizing that the tidal phase of particle release might influence the calculated residence time, we conducted three different simulations, by releasing passive particles at 02:00 (T1, low tide) and 17:00 (T2, high tide) on May 1 and 00:00 (T3, flood slack) on May 2, 2010, respectively. The water elevations at Grand Isle station for these three times were 0.027 m, 0.53 m, and 0.42 m, respectively. The calculated residence times are shown in Tab. 2.3. Obviously,
residence times had a strong spatial and temporal variability in this geometrically complex estuary. The results indicate that residence times are sensitive to releasing time. For instance, residence time of group 8, released at low tide (T1), was more than twice as long as that released at high tide (T2), in the DPD scenario. Both DPD and MBD had little impact on residence times of the upper estuary (group 1, 2,3,4, 9,10,12 and 19) and the lower estuary, which are located far away from the freshwater pathways (group 18). For the Barataria Waterway (groups 5 and 6), the MBD had a much greater impact on the residence time than the DPD, i.e., the maximum difference between the DPD and NO scenarios was 3 days, while the difference between the MBD and NO scenario was 39 days (group 6, T3) when avoiding circumstances where particles could move out of the estuary over the simulation period. The MBD also could impact the residence time of Little Lake (group 7), which was 50 days in the DPD scenario, but 15 days in the MBD scenario at T2. The adjacent area of the MBD (groups 14 and 15) was inundated by freshwater (Fig. 2.7f and 2.8f). The animation of particle transport shows that particles were transported through upstream tributaries to Barataria Waterway, thus making them easier for transport out of the estuary. Both DPD and MBD had significant impacts on the residence time of Barataria Bay (group 8). The residence time varied from 6 days at T3 to 44 days at T1 in the DPD scenario, while in the NO scenario this value varied from 14 days at T2 to 50 days at T1. In the MBD scenario, the residence time of Barataria Bay varied from 3 days at T2 to 18 days at T2. Based on 24-hour observations, Li et al. (2011) calculated the flushing time of the Barataria Bay to be between 13 to 19 days using an average DPD discharge of ~ 100 m$^3$ s$^{-1}$ over their study period. Our study used a discharge value of ~ 200 m$^3$ s$^{-1}$. This made 6 days for flushing in the DPD scenario reasonable.
Here residence time is defined as the time for a particle to leave Barataria Estuary once
and assign values of residence time to the locations of particle release. As stated by Monsen et al.
(2002) and Huang et al. (2011), this definition does not consider oscillating tidal transport of
water and scalars into and out of the estuary over multiple tide cycles. A different water
transport time scale, exposure time, which is defined as the total amount of time a particle spends
in the domain of interest, might be a more relevant time scale than residence time for some
geochemical or biological processes because particles that leave the domain can enter it again at
some later time. Therefore, the influence of estuary habitat could be greater on the particle than
would be indicated by the residence time calculated here.

Table 2.3. Residence time (days) for different releasing times (T1, T2, and T3) in the Barataria
Estuary under three diversion discharge scenarios. Note that 60.92 for T1, 60.29 for T2, and
60.00 for T3 means particles cannot move out of the estuary over simulation period.

<table>
<thead>
<tr>
<th>ID</th>
<th>DPD</th>
<th>NO</th>
<th>MBD</th>
</tr>
</thead>
<tbody>
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<td>T1</td>
<td>T2</td>
<td>T3</td>
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<td>60.29</td>
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</tr>
<tr>
<td>4</td>
<td>60.92</td>
<td>59.19</td>
<td>56.74</td>
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<tr>
<td>5</td>
<td>59.08</td>
<td>59.08</td>
<td>46.05</td>
</tr>
<tr>
<td>6</td>
<td>55.62</td>
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<td>43.72</td>
<td>17.15</td>
<td>5.55</td>
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<tr>
<td>8</td>
<td>60.92</td>
<td>60.29</td>
<td>60.00</td>
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</tr>
<tr>
<td>19</td>
<td>60.92</td>
<td>60.29</td>
<td>60.00</td>
</tr>
</tbody>
</table>
2.5.4. Water flux through transects along the Barataria Waterway

The water volume flux through different transects along the Barataria Waterway were used to investigate how diversions affect water transport. The water volume flux is defined as the amount of water volume that passes through the transect in unit time and is estimated using:

\[
Q = \int_{-H}^{\zeta} \left( \int_{0}^{L} u \, dx \right) \, dz
\]

where \( u \) is the normal velocity component, \( H \) is the water depth, \( \zeta \) is the water surface elevation, and \( L \) is the length of the transect. The magnitude of water flux depends on the area of the transect and normal velocity. Fig. 2.13 shows a time series of water flux at five transects (see Fig. 2.1 for locations) in the Barataria Waterway. All-time series show diurnal characteristics. Transect T1, T2, and T3 could be highly impacted by local wind effect, as can be seen on 24 April and 2 May, 2010. This is consistent with the results of Snedden (2006), in which cross-shelf winds were found to be a dominate factor forcing barotropic exchanges in the Barataria Bay. Snedden (2006) also found northward wind stress acting over the bay may result in setup at the northern end of the bay. The maximum water volume flux was at T5 (Fig. 2.13e). The flux can reach up to \( \sim 6000 \, m^3 \, s^{-1} \) and \( \sim 2000 \, m^3 \, s^{-1} \) during spring tides and neap tides, respectively. The net rate of flux for the MBD scenario increased by \( \sim 44\% \) compared with the DPD scenario. The water flux for transect T4 was relatively small, with magnitude less than 100 \( m^3 \, s^{-1} \) (Fig. 2.13d). This is because this transect is adjacent to the Barataria Bay. Even though it was still in the channel, upstream water spread out when it reached the bay. The MBD had a great impact on upstream water exchange (Fig. 2.13a, 2.13b, 2.13c). The net outward flux rate of the MBD scenario was up to 10 times as high as that of the DPD scenario (Fig. 2.13c). The residence time
upstream of the waterway also reflected the impact of the MBD, which was reduced to 4 days (Tab. 2.3 group 5).

Figure 2.13. Time series of water fluxes at five transects (shown in Fig. 2.2) along the Barataria Waterway over simulation period. The first 10 days of simulation were spin-up time and are not shown here. Values shown at the top represent three-month net rate of flux corresponding to each scenario. Negative signs denotes outflow.
2.5.5. Water flux through four tidal inlets

Marmer (1948) calculated the amount of water going through four passes, Barataria Pass (BP) (66%), Caminada Pass (CP) (13%), Pass Abel (PA) (3%), and Quatre Bayou Pass (QB) (18%), based on 24-day observations. His net rate of transport out of the system was ~ 280 m$^3$ s$^{-1}$. The three-month net flux in this study are shown in Tab. 2.4. For the NO scenario, the amount of water going through these four passes was 78% (BP), 11% (CP), 5% (PA), and 6% (QBP), respectively. For the DPD scenario they were 76%, 9%, 6%, and 9%. The MBD had little impact on this ratio, where the amount was 74%, 6%, 7%, and 13%, respectively. The net rate in DPD scenario is ~ 420 m$^3$ s$^{-1}$ out of the bay system through four passes. Without the DPD, the net rate out of the bay system was ~ 307 m$^3$ s$^{-1}$. In the MBD scenario, the net rate was about 986 m$^3$ s$^{-1}$ out of the system, which is almost twice as great as that of the DPD scenario. Fig. 2.14 shows a time series of water flux at the four passes. BP was the only one that had strong tidal variation. The other three passes were noise with more high frequency variations.

Table 2.4. Three-month net rate of water flux through four passes (BP – Barataria Pass, CP – Caminada Pass, PA – Pass Abel, QB – Quatre Bayou Pass, units in m$^3$ s$^{-1}$, negative values denote outflow).

<table>
<thead>
<tr>
<th></th>
<th>DPD (-420)</th>
<th>NO (-307)</th>
<th>MBD (-986)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BP</td>
<td>-320</td>
<td>-239</td>
<td>-726</td>
</tr>
<tr>
<td>CP</td>
<td>-40</td>
<td>-34</td>
<td>-63</td>
</tr>
<tr>
<td>PA</td>
<td>-24</td>
<td>-15</td>
<td>-70</td>
</tr>
<tr>
<td>QBP</td>
<td>-36</td>
<td>-19</td>
<td>-127</td>
</tr>
</tbody>
</table>

2.6. Conclusions

Sediment diversions have been considered a more efficient tool for rebuilding wetlands, compared to long-distance pipelines that spoil dredged materials from the river beds and inland to offshore deposits. The existing river diversion in the Barataria Estuary is Davis Pond Diversion, which was built to control salinity intrusion and flooding. A series of studies have
Figure 2.14. Time series of water fluxes of the (a) Barataria Pass, (b) Caminada Pass, (c) Pass Abel, and (d) Quatre Bayou. The positive and negative values denote flux in and out, respectively. Black line represents the DPD scenario, blue line represents the NO scenario, and red line represents the MBD scenario.

estimated the impact of river diversions on salinity gradients in Louisiana estuaries (Bianchi et al., 2011; Das et al., 2012; Inoue et al., 2008; Park et al., 2004). Das et al. (2012) found that the salinity difference can be as high as 10 in the middle section of the Barataria Estuary under differing discharge scenarios. However, the maximum discharge in their study was ~ 300 m$^3$ s$^{-1}$. The goal of the proposed Mid-Barataria Diversion is to provide sediment to the Barataria Basin, which can divert a maximum of ~ 2100 m$^3$ s$^{-1}$. The diversion would have a great impact on the
salinity gradient and residence time in the estuary, further affecting the diversity of fish species, as many species sensitive to salinity changes.

The model results show that the MBD, with a discharge of \( \sim 850 \text{ m}^3 \text{ s}^{-1} \) would have a significant impact on salinity in Barataria Bay, which is near the mouth of the estuary. The three-month average bottom salinity difference could be as high as 12 in the Barataria Bay, with an average surface salinity difference of \( \sim 4 \). The impact area from the MBD could extend from the middle estuary to lower estuary, almost half of the Barataria Estuary. The DPD model showed a minimal impact on average surface salinity but a potentially large impact on average bottom salinity of the Little Lake and Barataria Bay, with a salinity difference \( \sim 3 \). The residual currents within a radius of \( \sim 10 \text{ km} \) from the DPD could be greatly increased by diversion discharge. The surface residual velocity change was \( \sim 10 \text{ cm s}^{-1} \). The MBD could affect the residual currents within a radius of \( \sim 25 \text{ km} \) from the diversion structure. It also has an impact on residual velocities along the Barataria Waterway. The DPD may alter residence times of Little Lake and Barataria Bay but may have little impact on residence times in the upper portion. However, residence times of a large portion of the estuary could be affected by the MBD due to changes in residual currents and water exchanges.

The Mid-Barataria Diversion used in this model was chosen as a test bed for the analysis, and is not located at the exact proposed site. It can provide as cost-effective tool to guide future operation on sediment diversions.
CHAPTER 3. LATERAL CIRCULATION IN A PARTIALLY STRATIFIED TIDAL INLET

3.1. Introduction

The lateral circulation in estuaries can be driven by various mechanisms including those related to tides. Nunes and Simpson (1985) identified the effect of differential advection on the generation of secondary circulation. Theoretically, the along-channel velocity will be stronger over the channel than the shoals, resulting in greater (smaller) density at the thalweg than the shoals during flood (ebb) tides. This will produce a cross-channel pressure gradient toward the channel (shoals) on the surface and a pressure gradient toward the shoals (channel) at the bottom during flood (ebb) tides. Thus, the lateral flows are convergent (divergent) at the surface over the deep channel and divergent (convergent) at the bottom during flood (ebb) tides. However, the axial convergence is only observed during flood tides in Nunes and Simpson’s work. In an idealized, straight channel with weak stratification, Lerczak and Geyer (2004) also confirmed that secondary circulation was driven by differential advection. Differential rotation of tidal ellipse is also identified as a mechanism for axial convergence fronts (Li, 2002). Interactions between barotropic pressure gradient and bathymetry can generate convergence of lateral flow, producing flows rotating toward the channel from the shoals (Li and Valle-Levinson, 1999; 2000). In curved estuaries, an alternative driving mechanism for lateral circulation can be centrifugal acceleration (Chant and Wilson, 1997; Lacy and Monismith, 2001) and advection (Li et al., 2008). Ekman-forced lateral circulation varies with the Ekman number. When the boundary layer is comparable to the channel depth (large Ekman number), lateral flow is a single circulation cell; while for thin tidal boundary layer (small Ekman number), lateral flow is complex and varies over the tidal cycle (Lerczak and Geyer, 2004). Boundary mixing on a no-
flux boundary layer was confirmed to be one of the driving mechanisms of lateral circulation (Chen and Sanford, 2009; Lerczak and Geyer, 2004). Cheng et al. (2009) investigated the lateral circulation during stratified ebb tides due to the lateral baroclinic pressure gradient, which is generated by differential diffusion caused by a lateral asymmetry in vertical mixing.

Using an idealized, straight, stratified estuarine channel, Lerczak and Geyer (2004) found that lateral flow is about four times stronger during flood tides than during ebb tides. They attributed it to the interaction between the along-channel tidal currents and nonlinear advective processes over a tidal cycle. The flood-ebb asymmetry in the lateral circulation strength was also observed by Scully et al. (2009) in a realistic estuary, the stronger lateral flows observed during flood tides while suppressed lateral flows during ebb tides. However, in the numerical modeling of James River Estuary by Li et al. (2017), they found that during neap tides, lateral circulation is stronger during ebb tides than during flood tides. During spring tides, there is no flood-ebb asymmetry.

Lateral circulation may play an important role in estuarine dynamics. Many observational and numerical simulation results (Cheng et al., 2009; Geyer, 1993; Lacy and Monismith, 2001; Lerczak and Geyer, 2004; Vennell and Old, 2007) have demonstrated the existence of secondary currents, and discussed its dependence and feedback on stratification (Cheng et al., 2009; Lacy and Monismith, 2001; Lacy et al., 2003; Nidzieko et al., 2009), streamwise momentum budget (Lerczak and Geyer, 2004), and estuarine circulation (Cheng et al., 2009; Scully et al., 2009). In Barataria Estuary, observations (Li et al., 2009) show that there is an asymmetry in lateral stratification in the Barataria Pass. In this study, we investigate how the lateral currents vary in the Barataria Pass.
In this study, we use a 3-D, high-resolution hydrodynamic model to examine the lateral circulation in the partially stratified tidal inlet, Barataria Pass, which connects Barataria Bay with the coastal ocean in the southwestern Louisiana. The objectives of this study are to elucidate the tidal evolution of lateral circulation and determine its driving mechanisms. The remainder of this paper is structured as follows: Section 3.2 describes the study area, and the configuration of the three-dimensional finite volume numerical model. Section 3.3 presents the validation of the numerical model and the temporal evolution of lateral circulation over a tidal cycle. In section 3.4, we quantify the 2-D momentum balance and examine the driving forcing for lateral circulation with 3-D momentum equations. Flood-ebb variations in lateral circulation are also discussed in section 3.4. Conclusions are discussed in section 3.5.

Figure 3.1. (a) Geographic location of the Barataria Estuary. S1-S6 are USGS stations. (b) Location of Barataria Pass. The coordinate is defined positive $x$ to the eastern bank, positive $y$ to the upstream. The black line indicates the cross-section used in this study. (c) The vertical cross-section. The black lines indicate CTD measurements. C1-C7 are locations used for 2-D momentum equation analysis.
3.2. Materials and methods

3.2.1. Study area

Barataria Bay is located at the southeastern Louisiana, on the western side of the Mississippi Birdfoot Delta. It is connected to the gulf by several tidal inlets including Barataria Pass (BP), which is between two barrier islands, Grand Isle Island and Grand Terre Island (Fig. 3.1b). The main freshwater comes from Davis Pond freshwater diversion, runoff and precipitation. The maximum diversion discharge is up to 300 m$^3$ s$^{-1}$ (Das et al., 2012). Barataria Pass is an 800 m wide narrow channel. It is one of the four main tidal passes of Barataria Basin, accounting for ~ 66% of total water exchange (Marmer, 1948). Tidally induced flow accounts for ~ 85% of the total variability, with equal contributions from the $O_1$ and $K_1$ constituents. Maximum tidal magnitude is relatively small, which is ~ 0.3 m. Barataria Bay is composed of broad shallow waters (average depth of 2 m), islands and a 5 m deep main shipping channel, the Barataria Waterway. The shipping channel ends at Barataria Pass. The main inlet has an average depth of ~ 20 m and is being periodically dredged, causing a depression of ~ 50 m deep close to the inlet (Fig. 3.2).

3.2.2. Model description and configuration

In this study, we use the Finite Volume Coastal Ocean Model (FVCOM) to simulate the hydrodynamics of the Barataria Basin and adjacent continental shelf. FVCOM is a three-dimensional, hydrostatic, free surface, primitive-equation ocean model (Chen et al., 2011; Chen et al., 2003). In the finite volume method, the computational domain is discretized using a mesh of non-overlapping triangles in the horizontal and sigma-coordinate in the vertical. The governing equations are solved in their integral forms in individual control volumes. The
Figure 3.2. Unstructured grid configured for the FVCOM Barataria Pass model: (a) regional domain; (b) local domain of Barataria Estuary; (c) local domain of Barataria Pass, with horizontal resolution ~50 m in the cross-channel direction and 30 m in the along-channel direction. Contours are interpolated bathymetry.

Triangular grid in the horizontal can resolve complex coastal and bathymetric geometries. It uses a cell-vertex-centered (similar to the finite-difference C-grid) method, which facilitates the enforcement of mass conservation in tracer advection and tracer open boundary conditions.

Vertical mixing uses modified Mellor and Yamada level 2.5 turbulence model (Galperin et al., 1988; Mellor and Yamada, 1982) and horizontal diffusion uses Smagorinsky eddy parameterization (Smagorinsky, 1963). The model employs mode split approaches barotropic 2D external mode and baroclinic 3D internal mode to solve the momentum equations with second-
order accuracy. The bottom boundary conditions apply an exact form of the no flux boundary conditions. Flooding/drying scheme is implemented in FVCOM. If vertical water column thickness at the cell center is less than a criterion value (typically 5 cm), then the cell is designated as a dry cell and its velocity is set to zero. Whenever the vertical water column thickness exceeds the criterion value, the cell becomes wet again and water level and velocity are computed from control equations. The advantage of triangular mesh to accurately represent complex bathymetry and coastlines makes FVCOM ideally suited for Barataria Pass study.

A high-resolution FVCOM Barataria Pass model was developed by configuring FVCOM version 2.6 to the Northwestern Gulf of Mexico continental shelf region with inclusion of the intertidal zones inside the Barataria Bay. The computational domain extends longitudinally from Mobile Bay, Alabama to west of Galveston Bay, Texas and offshore to about 27°N (Fig. 3.2). It consists of 146,266 triangular nodes and 283,721 triangular cells. The horizontal grid resolution varies from about 10 m in the upper estuary to about 8 km near the open boundary. Near the Barataria Pass the grid cells are fine enough to ensure that the 800 m wide channel cross section is resolved by ~ 20 triangles (Fig. 3.2c). The vertical resolution is 19 uniform sigma layers, which is ~ 0.1 m over the shoal and ~ 1 m in the central depression of the tidal inlet. Computational time steps are 0.2 s and 2 s for the external and internal modes, respectively. Model results are saved every 10 minutes for further diagnostic analysis.

Model bathymetry is obtained from various sources. Using an inverse distance method, a 5 m by 5 m resolution digital elevation model (DEM) constructed from Light Detection and Ranging (LIDAR) dataset is interpolated into model wetland region. The water depth in channels, bayous, and lakes is interpolated from Coastal Louisiana Ecosystem Assessment and Restoration Report (CLEAR), US Army Corp Survey, and NOAA nautical charts. Shelf and open ocean
water depth is interpolated from a coarse resolution ADCIRC model bathymetry. The water depth values in the inlet were obtained with vessel-based surveys (Li et al., 2009; Li et al., 2011).

Salinity is considered to be the most important factor that influences density and vertical stratification in most estuaries and in the Barataria Bay and adjacent coastal oceans. Thus, it is set to be the only prognostic tracer variable in FVCOM simulations. Temperature is kept as a spatial and temporal constant (20°C). The coefficients for horizontal dissipation and diffusion are both set to be 0.4 m² s⁻¹. The conventional quadratic bottom friction formulation is applied, with drag coefficient C_d determined by matching a logarithmic bottom boundary layer velocity to that of the numerical model at the lowest sigma-layer height. However, bottom drag coefficient over the wetlands is defined as five times greater than that in the estuarine channels, mimicking the vegetation damping effect (Huang et al., 2011).

3.2.3. Model forcing initial and boundary conditions

The model is driven by winds at the surface, sea level elevation at the open boundary, and freshwater inflows from Mississippi River passes, various Atchafalaya River passes, and the Davis Pond Diversion. It is initialized on 1 October 2007 and run until 31 December 2008. We use 3-hourly wind data from NOAA National Centers for Environmental Prediction (NCEP) North American Regional Reanalysis (NARR) products and interpolate it into the entire computational domain. The initial values of sea level elevation and velocity are specified as zero throughout the computational domain. The initial salinity field over the continental shelf is interpolated from HYCOM Gulf of Mexico 1/25° reanalysis product, while salinity inside the Barataria Bay is interpolated from observations at the USGS stations. Experiments show that this technique gives more accurate salinity simulation result than that with piecewise linear
interpolation from estuarine head to the mouth. At the open boundary, the 6-min interval sea
level time series at four stations are downloaded from NOAA tides and currents website. Time
series of Dauphin Island, Southwest Pass, Freeport, and Galveston Pier 21 are directly used to
prescribe sea level at the easternmost node, the southeastern node, the southwestern node, and
the westernmost node, respectively. Sea level elevations at other open boundary nodes are
linearly interpolated from these four nodes. Observed 15-min freshwater discharge at four
locations, Mississippi River at Belle Chasse, David Pond diversion, Atchafalaya River at Morgan
City, and Wax Lake Outlet, are injected into the computational domain with flux boundary
conditions of zero salinity and specified volume and momentum. All model forcing functions are
ramped up from zero over a period of 10 days.

3.2.4. Observations

Simulated water elevation, velocity and salinity are compared with in situ observations.
Water elevation data is obtained from National Water Information System of USGS shown in
Fig. 1, including six stations, Barataria Bay near Grand Terre Island (S1), Barataria Bay North of
Grand Isle (S2), Hackberry Bay near NW of Grand Isle (S3), Barataria Waterway (S4), Little
Lake near Bay Dosgris (S5), and Little Lake near Bay Cutoff (S6). The unit of water elevation is
converted to meter and vertical datum is adjusted to mean sea level in order to be consistent with
model results. For this study we focus on the time period from 00:00 31 July to 00:00 2 August
2008 in order to compare with observations. Velocity and salinity data are from the field
observations conducted at Barataria Pass from 11:30 am July 31 to 11:10 am August 1, 2008
UTC (See Li et al. (2009) and Li et al. (2011) for details of data description).
3.3. Results

3.3.1. Two-month water elevation comparisons

We use the cross-correlation coefficient to estimate the performance of the model. Two-month (1 July to 31 August, 2008) time series of water elevation for both observation and model simulation are shown in Fig. 3.3. The model reproduces the observed tidal variations, including tropic-equatorial modulation and surface set-up, with all correlation coefficients above 0.9.

3.3.2. Velocity

Similar to the treatment in Li et al. (2009), the coordinate system was counterclockwise rotated by 52.7° from the east-west direction to obtain the cross- and along-channel velocity components, which are shown in Fig. 3.4 Positive is flood current. As shown in Fig. 3.4a, the observed along-channel velocity has a stronger tidal signal, with maximum magnitude of ~ 1.5 m s⁻¹, than the model simulated velocity, which has a maximum magnitude of ~ 1.0 m s⁻¹. The tidal phase was in agreement with the observations. Both observed and modeled cross-channel velocities (Fig. 3.4b) are much smaller and noisier, and the tidal signal is not clear, compared to the along channel velocity component.

3.3.3. Vertical salinity profile

A total of 28 CTD casts were made between July 31 and Aug. 1, 2008 during a 25.6 hour period (See Tab.1 in Li et al. (2011)for details). Vertical salinity profiles at three locations, eastern, middle, and western side of the channel (Fig. 3.5), is compared with the CTD measurements. The magnitude of observed salinity ranges between 19 and 28.5. The maximum vertical salinity difference is about 5.5. The magnitude of the simulated salinity varies between
15 and 27. Generally, the model underestimates salinity. This is likely because we do not include evaporation and precipitation in the simulation. However, the model successfully captures characteristic features in salinity vertical profile. For example, cast 6 (Fig. 3.5a), which was made 4 hours after maximum flood, has a weak stratification at the top of the water column and well-mixed state at the bottom. Our model captures this feature. Other casts, such as casts 19, 26, 18, 8, and 16, have similar vertical profiles with the simulation. The temporal evolution of modeled salinity is also consistent with the observations. For example, the sequence of cast numbers (from low to high) for both the observation and the model show that the salinity tends to decrease in Fig. 3.5f (ebb tide, west station). This gives us confidence to apply the model to do further physical dynamic analysis.

3.3.4. Temporal variation of stratification in the Barataria Pass

In order to study the variation of stratification during a diurnal cycle, a time duration of ~25.6 hours from the FVCOM simulation, starting from 06:00 31 July to 07:40 1 August, 2008, was chosen to do further analysis. Time series of water level, depth-averaged along-channel velocity and salinity difference between bottom and surface at the three locations, are shown in Fig. 3.6. The water elevation and depth-averaged velocity are not in-phase, with phase difference ~60°, indicating a standing wave behavior. Stratification at these three locations varies. Within 2 hours of early flood tides, stratification at all three locations decreases and reaches a well-mixed condition. Then stratification starts and increases. But the station on the west quickly reaches the maximum (Fig. 3.6a), followed by the station at the middle (Fig. 3.6b). During the remaining period of flood tides, the western shoal experiences variation between well-mixed and stratified conditions. The middle channel and the eastern shoal are always stratified. During ebb tides, stratification at the western shoal decreases in the beginning, and the western shoal remains well-
mixed almost over the whole ebb tides. Stratification in the middle and that in the eastern shoal share the same characteristics - decreasing in the beginning and then fluctuating around zero for a long time, before increasing again at the end.

Figure 3.3. Water elevation comparison between USGS observations (black) and model simulation (red) from July 1 to August 31, 2008. Grayed areas represent missing data.
Figure 3.4. (a) Along-channel velocity at 1.32 m below the surface for observed (red dot) and modeled (blue dot); (b) Cross-channel velocity at 1.32 m below the surface for observed (red dot) and modeled (blue dot).

Figure 3.5. Vertical profile of salinity comparison between in situ observation (red) and simulation (black). The numbers in the plot represent the sequence of CTD casts. Left panels are during flood tides. Right panels are during ebb tides. Top row is at the eastern side of the channel, middle row is at the middle of the channel, and bottom row is at the western side of the channel.
Figure 3.6. Time series (06:00 31 July to 07:40 1 August, 2008) of water level elevation (red), depth-averaged along-channel velocity (green), and bottom-top salinity difference at (a) western shoal, (b) deep channel, and (c) eastern shoal. Shaded area represents flood tide. Arrows show different stages during a tidal cycle.
3.3.5. Residual currents over one tidal cycle

Tidal, ebb-, and flood- averaged along-channel velocities are shown in Fig. 3.7. All the plots show that the western bank is on the left and the eastern bank is on the right. The magnitude of spatially averaged ebb tides is -46 cm/s, and that of spatially averaged flood tides is 33 cm/s. The transverse structure of the along-channel residual current differs significantly between ebb and flood tides. During ebb tides, the maximum outflow is at the surface near the western shoal, inclining against the western bank. During flood tides, the maximum inflow is near the bottom in the central deep channel. The magnitude of spatially averaged residual current during the 25.6-hour period is -9 cm s⁻¹, which is in the ebb direction. The maximum residual current is near the western shoal, with the magnitude up to -30 cm/s. With idealized experiments, Cheng et al. (2009) addressed that the exchange flow pattern sensitively depends on the nondimensional Rossby number \( R_0 \) (\( U/fB \), where \( U \) is the estuarine circulation velocity, \( f \) is the Coriolis parameter, and \( B \) is the width of the channel) and Ekman number \( E_k \) (\( A_z/fH^2 \), where \( A_z \) is the vertical eddy viscosity, \( H \) is water depth, and \( f \) is the Coriolis parameter). They demonstrated the exchange flow is vertically sheared at large \( R_0 \), and horizontally sheared at large \( E_k \). In our case (\( R_0 \approx 1.72, E_k \approx 2.06 \)), the residual current is both vertically and horizontally sheared, which is consistent with idealized cases Fig. 5c (\( R_0 = 2.63, E_k = 1 \)) of (Cheng and Valle-Levinson, 2009). For a triangular shaped cross section, Wong (1994) showed that the estuarine circulation is outward at the surface and inward at the bottom of the deep channel due to the interaction between baroclinic force and triangular bathymetry. The observations capture this characteristics (Li et al., 2011), but the inflow is very weak. Our model also reproduces a weak inflow. However, this inflow occurs near the bottom of the eastern slope (Fig. 3.7c), which may be caused by the reverse tidal asymmetry (Cheng et al., 2010), that is
weaker mixing during flood than during ebb. The asymmetry of estuarine-ocean exchange, i.e., inflow tending to the right side of the channel and outflow to the left side (when looking up-estuary) can be attributed to the Coriolis force (Geyer et al., 1997).

The tidally-averaged cross-channel velocity is weak (Fig. 3.8), with a magnitude less than 10 cm/s. The magnitudes of flood-averaged and ebb-averaged velocity can reach 20 cm/s. The transverse structures of flood-averaged and ebb-averaged velocities are different. The ebb-averaged velocity exhibits a horizontal gradient, with a weak divergent region inclining to the western slope (Fig. 3.8e), while the flood-averaged velocity shows a vertical gradient, with a convergent region inclining to the eastern slope (Fig. 3-8d). Tidal variation of axial convergence front is anticipated (Li, 2002). Tidally-averaged lateral velocity displays a clockwise, single-cell circulation (Fig. 3.8f).

3.3.6. Cross-sectional salinity and currents structures over a tidal cycle

Fig. 3.9 shows the temporal evolution of along-channel velocity, lateral circulation, salinity, and turbulent vertical eddy viscosity for the Barataria Pass transect during flood tide. The individual time instance for each row is shown in Fig. 3.6 as a vertical arrow. One hour after the flood starts (T1 in Fig. 3.6), stratification is weak (bottom-top salinity difference ~ 1) across the deep channel and eastern shoal, and the western shoal is almost well-mixed (Fig. 3.9c). Strong lateral circulation mainly occurs in the mid-layer of the channel with a convergence zone below 10 m in the west channel (Fig. 3.9b). The along-channel velocity is ~ 0.4 m/s, extending almost the whole deep channel. Thus, vertical and lateral velocity shear is weak. The vertical eddy viscosity is mostly smaller than 0.005 m²/s, probably due to the small flood current magnitude.
Figure 3.7. Transverse distribution of ebb-, flood-, and tidally-averaged along-channel velocities, looking up-estuary (unit: cm/s)

Figure 3.8. Transverse distribution of ebb-, flood-, and tidally-averaged cross-channel velocities (unit: cm/s, positive is eastward, negative is westward)

Two hours later (T2 in Fig. 3.6), the whole water column becomes more or less well-mixed across the deep channel and eastern shoal, while in the western shoal a sharp salinity stratification develops with bottom-top salinity difference ~3 in 2 m water column (Fig. 3.9g). The distribution of salinity, and thus density, in the cross-channel direction is such that lateral
baroclinic pressure gradients are directed from the central part of the deep channel towards the shoals. This is similar to the situation pointed out in Nunes and Simpson (1985). Based on their theory, such pressure gradient force will induce a lateral circulation with convergence at the surface and divergence near the bottom. This indeed occurs in our numerical results. Surface lateral flow at the west half of the channel changes from ~0.1 m/s westward to ~0.2 m/s eastward between T1 and T2. A pair of counter-rotating circulations is clearly seen at T2 with strong convergence at 2 m below the surface. Contrast to the idealized case in Lerczak and Geyer (2004), the two circulation cells are not closed at this time. The maximum along-channel tidal velocity reaches ~0.8 m/s, confined at the mid-depth (Fig. 3.9e). The vertical shear in along-channel velocity is weak, while the horizontal shear is great, which is, over the western slope, ~0.6 m/s within 100 m distance. It seems Nunes and Simpson’s argument is also applicable here in that lateral differential advection is at least one of the mechanisms to generate the lateral salinity gradient. Strong vertical mixing (maximum eddy viscosity ~0.05 m²/s, Fig. 3.9h) occurs at the mid-depth and bottom boundary layer, where either tidal currents or bottom friction are strong (Fig. 3.9e). Strong turbulence mixing tends to destratify the water column, which explains the relatively uniform salinity distribution in the deep channel and east shoal (Fig. 3.9g).

At the maximum flood (T3), the along-channel velocities intensify. Maximum along-channel velocity reaches ~1.2 m/s and extends to the surface (Fig. 3.9i), which is quite different from T2 (Fig. 3.9e). The lateral shear of along-channel velocity, i.e. differential advection, is 1.0 m/s across 300 m distance on both sides. However, salinity distribution changes drastically. The western shoal is completely well-mixed at this time, while the eastern shoal and part of the east channel have surface stratification with a salinity difference of ~3 within 5-6 m depth (Fig. 3.9k). The stratification in the deep channel, especially below 10 m, is weak, because the tidal mixing
is relatively high (vertical eddy viscosity ~0.03 m$^2$/s) at the bottom boundary (Fig. 3.9l). The lateral circulation pattern is similar to that at T2, but with intensified strength. The convergent zone rises to the ocean surface and the right circulation cell is now a complete circle (Fig. 3.9j).

Four hours before the end of flood (T4), differential advection still persists, although the maximum along-channel tidal current has reduced to ~1 m/s and located below the surface (Fig. 3.9m). Turbulence mixing weakens (Fig. 3.9p). Hence, a weak stratification (bottom-surface salinity difference ~1.5) develops at the western shoal (Fig. 3.9o). The salinity distribution shows a more symmetric pattern relative to the axis of the channel compared with T1, T2, and T3. As a result, the counter-rotating lateral circulation cells are more symmetric and fully developed (Fig. 3.9n).

At the flood slack (T5), the surface water column becomes stratified in the upper 6 m (salinity difference ~ 1) while the deep channel is almost well-mixed (Fig. 3.9s). The lateral circulation almost completely disappears in the deep channel (Fig. 3.9r).

At the beginning of ebb tide (T6), although the along-channel ebb current increases to ~0.4 m/s (Fig. 3.10a), the salinity distribution and vertical stratification (Fig. 3.10c) are almost the same as that of T5. There is a weak (less than 0.1 m/s) eastward lateral flow below 6 m (Fig. 3.10 b).

Two hours later (T7), the along-channel ebb current reaches above 1.0 m/s, locating mostly on the western slope. The magnitude of along-channel velocity across the majority of the cross section is ~ 0.8 m/s (Fig. 3.10e), which induces largest tidal mixing (maximum vertical eddy viscosity ~ 0.16 m$^2$/s, Fig. 3.10h). Thus, the whole water column is vertically well-mixed. But a horizontal salinity gradient exists, with higher salinity located near the channel axis,
fresher water on both shoals. The western shoal is fresher than the eastern shoal (Fig. 3.10g). This is because freshwater is flushed out of the estuary through the western shoal, as shown in Figs. 3.7b and 3.7c. The lateral circulation shows mostly eastward currents in the deep channel across the whole water column, while the eastern shoal has a convergent area (Fig. 3.10f).

During the next five hours, this cross section is always vertically well-mixed and salinity decreases constantly due to freshwater outflow. The turbulence mixing remains intense in the deep channel during this period. The maximum vertical eddy viscosity can reach up to 0.2 m²/s, which results in the water column in the east half of the channel vertically and horizontally well-mixed (Fig. 3.10k, T8). The water column in the west half of the channel is also vertically well-mixed, but has a weak (~ 1) horizontal salinity decrease westward. The structure of lateral circulation and along-channel velocity remains the same, although the maximum ebb velocity has decreased from 1.0 m/s at T7 to 0.8 m/s at T8.

Later during the ebb period (T9), the vertical stratification returns (Fig. 3.10o). The vertical salinity difference is ~ 2 in the deep channel and on the eastern shoal. Lateral circulation shows a weak flow divergence close to the surface near the western slope (Fig. 3.10n). The maximum along-channel ebb velocity is located at the surface and decreases to 0.6 m/s (Fig. 3.10m), while turbulence vertical eddy viscosity decreases greatly (Fig. 3.10p).

At the ebb slack (T10), salinity distribution go back to similar to T1 situation. The western shoal is almost vertically uniform. Strong stratification (salinity difference ~2 over 6-m depth) occurs in the upper water column of the deep channel and on the eastern shoal. The lower water column in the deep channel has a weak stratification (Fig. 3.10s). Lateral circulation is greatly reduced compared to other time instances of the ebb tide (Fig. 3.10r).
Figure 3.9. Cross-sectional profiles of currents ($u, v, w$), salinity and vertical viscosity during flood tide. The first column is along-channel velocity, the second column secondary circulation, the third column salinity, and the last column vertical viscosity. The velocity contours are 0.2 m/s, positive is up-estuary. The salinity contours are 0.5.
Figure 3.10. Cross-sectional profiles of currents ($u$, $v$, $w$), salinity and vertical viscosity during ebb tide. The first column is along-channel velocity, the second column secondary circulation, the third column salinity, and the last column vertical viscosity. The velocity contours are 0.2 m/s, positive is landward. The salinity contours are 0.5.
3.4. Discussion

3.4.1. Depth-averaged momentum balance

The vertically averaged cross- and along-channel momentum equations are written as:

\[
\frac{1}{D} \frac{\partial \bar{u} D}{\partial t} = -\frac{1}{D} \left( \frac{\partial \bar{u}^2 D}{\partial x} + \frac{\partial \bar{u} \bar{v} D}{\partial y} \right) + f \bar{v} - g \frac{\partial \zeta}{\partial x} - \frac{g}{\rho_0} \int_{-1}^{0} \left( D \int_{\sigma}^{0} \rho d\sigma' \right) d\sigma + \frac{\partial D}{\partial x} \int_{-1}^{0} \sigma \rho d\sigma
\]

\[+ \frac{\tau_{sx}}{D \rho_0} \frac{\tau_{bx}}{D \rho_0} + \bar{F}_x + \frac{1}{D} G_x \]  

(3-1)

\[
\frac{1}{D} \frac{\partial \bar{v} D}{\partial t} = -\frac{1}{D} \left( \frac{\partial \bar{u} \bar{v} D}{\partial x} + \frac{\partial \bar{v}^2 D}{\partial y} \right) - f \bar{u} - g \frac{\partial \zeta}{\partial y} - \frac{g}{\rho_0} \int_{-1}^{0} \left( D \int_{\sigma}^{0} \rho d\sigma' \right) d\sigma + \frac{\partial D}{\partial y} \int_{-1}^{0} \sigma \rho d\sigma
\]

\[+ \frac{\tau_{sx}}{D \rho_0} \frac{\tau_{by}}{D \rho_0} + \bar{F}_y + \frac{1}{D} G_y \]  

(3-2)

where \((\bar{u}, \bar{v})\) are the vertical integrated cross- and along-axis velocity components. The positive \(u\) is pointed to the eastern bank, the positive \(v\) to the upstream. Terms from the left to the right are local acceleration (DDT), nonlinear advection (ADV), Coriolis force, barotropic pressure gradient (DPBP), baroclinic pressure gradient (DPBC), wind stress (WIND), bottom friction (FRIC), horizontal diffusion (HDIF), and difference between nonlinear terms of vertically-averaged 2-D variables and vertical integration of 3-D variables (AV2D). The expressions for the horizontal diffusion and AV2D can be referred to Chen et al. (2011). Consistent with currents converted to cross- and along-channel directions, all terms in eqs (3-1) and (3-2) are rotated from FVCOM x-y coordinate.

The Coriolis force, wind stress, and horizontal diffusion are at least one order of
magnitude smaller than the other terms. Thus these three terms are neglected. Fig. 3.11 shows
time series of six terms (DDT, ADV, DPBP, DPBC, FRIC, and AV2D) in Eqs. (3-1) and (3-2) at
7 locations across the channel (Fig. 3.1c). The momentum balance across this narrow channel is
much more complex, as various locations have different characteristics.

In the along-channel momentum balance, the dominant balance is between the barotropic
pressure gradient and the nonlinear advection, especially during ebb tides. The magnitudes of
these two terms for stations on the western side (C2 and C3, Fig. 3.11i, j) of the channel is at
least twice greater than that of other stations. This is because maximum ebb currents flush out
near these two stations. The sign of the barotropic pressure and nonlinear advection of stations
on the left side (C1-C3, Fig. 3.11h-j) of the channel is opposite to those on the right side (C4-C7,
Fig. 3.11k-n) of the channel, which is also related to the ebb currents. There is a spike during ebb
tides, similar to that in Huang et al. (2011). During flood tides, the magnitudes of all terms of
stations on the western side (C1-C3, Fig.3.11h-j) are relatively small. This is because the sign of
upper layer is opposite to that of the lower layer during flood tides. They offset each other after
integrating over depth. The balance is among the DDT, nonlinear advection, barotropic pressure
gradient, baroclinic pressure gradient, and the AV2D.

In the cross-channel momentum balance, the characteristics are similar to that of along-
channel, i.e., the dominant balance is between advection term and the barotropic pressure
gradient. Except station C1, the signs of advection and barotropic pressure at the westmost
station (C2, Fig. 3.11b) and the eastmost station (C7, Fig. 3.11g) are the same, but opposite to the
stations of the deep channel (C3, C4, Fig. 3.11c, d). During flood tides, the balance is among the
DDT, nonlinear advection, barotropic pressure gradient, baroclinic pressure gradient, and the
Figure 3.11. Time series of vertically averaged terms in horizontal momentum equations in the across- (a-g) and along-channel (h-n) directions during a tidal cycle. The left column is for the cross-channel direction and the right column the along-channel direction. DDT (dash black) represents the local acceleration, AVD (red) the non-linear advection, COR (pink) the Coriolis force, DPBP (green) the barotropic pressure gradient, DPBC (blue) the baroclinic pressure gradient, AV2D (purple) the difference between 2-D and 3-D nonlinear terms, FRIC (orange) the bottom friction, and HDIF (yellow) the horizontal diffusion. Shaded areas indicate flood tide. Stations (C1-C7) from top to bottom are located from the west to the east shown in Fig. 3.1c. Note that the y-axis scales for Fig. i and j are different from others.
AV2D. Note that, the baroclinic pressure is great at stations at the deep channel, and the magnitude is larger than that of along-channel. This indicates the lateral salinity gradient may play an important role in momentum balance.

3.4.2. Driving mechanism of lateral circulation

Lerczak and Geyer (2004) pointed out that lateral advection plays an important role in the estuarine dynamics when lateral flows are strong enough to advect water parcels relative to 0.5 times the breadth of the channel \(4(|v|)/\sigma B \geq 1\), where \(|v|\) is the absolute value of lateral velocity amplitude, \(\sigma\) is the semidiurnal tidal frequency, and \(B\) is the channel width) over a tidal cycle. As shown in previous results, lateral circulation in the Barataria Pass is strong both during maximum flood and maximum ebb. Thus, lateral advection is expected to be an important term in the momentum balance. Three-D depth-dependent \(\sigma\)-coordinate momentum equations were used to identify the mechanisms that drive lateral circulation. The equation is written as:

\[
\frac{1}{D} \frac{\partial u_D}{\partial t} = -\frac{1}{D} \frac{\partial u^2_D}{\partial x} - \frac{1}{D} \frac{\partial uD}{\partial y} - \frac{1}{D} \frac{\partial u\omega}{\partial \sigma} + fu - g \frac{\partial \zeta}{\partial x} - \frac{g}{\rho_0} \left[ \frac{\partial}{\partial x} \left( D f \int_0^\sigma d\sigma' + \sigma \rho \frac{\partial D}{\partial x} \right) \right] + \frac{1}{D^2} \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial u}{\partial \sigma} \right) + F_x
\]

\[
\frac{1}{D} \frac{\partial v_D}{\partial t} = -\frac{1}{D} \frac{\partial uD}{\partial x} - \frac{1}{D} \frac{\partial v^2_D}{\partial y} - \frac{1}{D} \frac{\partial v\omega}{\partial \sigma} - fu - g \frac{\partial \zeta}{\partial y} - \frac{g}{\rho_0} \left[ \frac{\partial}{\partial y} \left( D f \int_0^\sigma d\sigma' + \sigma \rho \frac{\partial D}{\partial y} \right) \right] + \frac{1}{D^2} \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial v}{\partial \sigma} \right) + F_y
\]

where \((u, v)\) are cross- and along-channel velocity components. The advection terms are moved to the same side of the pressure gradient, and each term in eqs. (3-3) and (3-4) is calculated with its corresponding sign (e.g., \(-\partial(u^2D)/D\partial x\)). Transverse distributions of momentum terms of
along- and cross-channel at T3 (flood tides) are shown in Fig.3.12 and Fig.3.13, respectively. In the along-channel momentum equation, baroclinic pressure gradient (Fig. 3.12e), Coriolis force (Fig.3.12 h) and stress divergence (Fig. 3.12i) are at least two orders of magnitude less than other terms at this time, thus are less important in the momentum balance. The total pressure gradient comes from the barotropic pressure gradient, but smaller than advection terms. Lateral advection of along-channel momentum (Fig. 3.12b, $-uv$) is great, playing an important role in the momentum balance. This term is mainly balanced by the along-channel advection ($-vv$). In the cross-channel balance, baroclinic pressure gradient is clearly evident at the bottom of the deep channel (Fig. 3.13e), with positive over the channel and negative at the shoal. This is consistent with differential advection induced cross-channel baroclinic pressure gradient that salinity is higher in the channel than at the shoals. Stress divergence is confined to the near bottom of the eastern slope. Differential advection is driving force of axial convergence during flood tides, which is consistent with previous study.

During ebb tides, the driving force of lateral circulation is different. Transverse distributions of momentum terms of along- and cross-channel at T7 (ebb tides) are shown in Fig.3.14 and Fig. 3.15, respectively. Lateral advection (Fig. 3.14b) of along-channel momentum is relatively smaller than along-channel advection (Fig. 3.14c), especially on the western side. The cross-channel baroclinic gradient (Fig. 3.15e) is small. The dominant cross-channel pressure gradient (Fig. 3.15f) is barotropic pressure gradient (Fig. 3.15d). Both lateral advection and pressure gradient generate eastward flow, which are balanced by along-channel advection (Fig. 3.15c). Except for lateral advection, barotropic pressure gradient also acts as driving force of lateral circulation during ebb tides.
Figure 3.12. Transverse distributions of all terms in the along-channel momentum equation at flood (T3). The scale is $10^{-3}$. Terms include (a) local acceleration, (b) lateral advection, (c) along-channel advection, (d) barotropic pressure gradient, (e) baroclinic pressure gradient, (f) total pressure gradient, (g) horizontal stress divergence, (h) Coriolis force, and (i) vertical stress divergence. The contour intervals are shown in parenthesis.
Figure 3.13. Transverse distributions of all terms in the cross-channel momentum equation at flood (T3). The scale is $10^{-3}$. Terms include (a) local acceleration, (b) lateral advection, (c) along-channel advection, (d) barotropic pressure gradient, (e) baroclinic pressure gradient, (f) total pressure gradient, (g) horizontal stress divergence, (h) Coriolis force, and (i) vertical stress divergence. The contour intervals are shown in parenthesis.
Figure 3.14. Transverse distributions of all terms in the along-channel momentum equation at ebb (T7). The scale is $10^{-3}$. Terms include (a) local acceleration, (b) lateral advection, (c) along-channel advection, (d) barotropic pressure gradient, (e) baroclinic pressure gradient, (f) total pressure gradient, (g) horizontal stress divergence, (h) Coriolis force, and (i) vertical stress divergence. The contour intervals are shown in parenthesis.
Figure 3.15. Transverse distributions of all terms in the cross-channel momentum equation at ebb (T7). The scale is $10^{-3}$. Terms include (a) local acceleration, (b) lateral advection, (c) along-channel advection, (d) barotropic pressure gradient, (e) baroclinic pressure gradient, (f) total pressure gradient, (g) horizontal stress divergence, (h) Coriolis force, and (i) vertical stress divergence. The contour intervals are shown in parenthesis.
3.4.3. Flood-ebb asymmetry

Here, we follow Lerczak and Geyer (2004) using the cross-channel average of depth-averaged velocity amplitude $<|u|> = \frac{1}{A} \iint |u|dA$ ($|u|$ is the absolute value of depth-averaged cross-channel velocity, $L$ is the width of the channel) to represent the strength of the lateral flow, which is shown in Fig. 3.16. The maximum $<|u|>$ at flood is 32 cm s$^{-1}$, while the maximum at ebb is 34 cm s$^{-1}$. Generally, cross-channel current amplitude during ebb is comparable, even a slight greater than that during flood, which is inconsistent with idealized case (Lerczak and Geyer, 2004), where lateral circulation is about 4 times as strong during flood than that during ebb. However, this inconsistence is also observed in James River estuary (Li, Liu, et al., 2017), where the lateral circulation shows no flood-ebb asymmetry during spring tides, and a reversed asymmetry during neap tides, that is stronger during ebb than during flood. Li et al. attribute negligible flood-ebb variations during spring tides to turbulence mixing, which reduces the vertical shear and the flood-ebb asymmetry in the vorticity generation. In the idealized case of Lerczak and Geyer (2004), they use constant eddy coefficients. This may be the reason for flood-ebb asymmetry. In our case, turbulence mixing is twice as great during ebb tides than during flood tides, which cause along-channel velocity during ebb tides almost vertically uniform.

During flood tides, lateral circulation shows asymmetry across the section (Fig. 3.9f, j, and n). The counterclockwise circulation on the right side (when looking landward) was stronger. The reason of increased circulation on the right side is due to the lower salinity water from the coastal ocean (Li et al., 2011), which enhanced lateral advection on the right side. The low salinity water is from the Mississippi River plume. During ebb tide, the single clockwise circulation cell lasted almost over the whole period. Only at the later ebb, there was a weak
counterclockwise circulation on the left side. This is because the along channel velocity moved from the western slope to the middle of the channel (Fig. 3.10i, m).

Figure 3.16. Sectional average of lateral velocity amplitude.

3.5. Conclusions

Barataria Pass is a tidal inlet that connects the Barataria Bay to the continental shelf. Previous investigation has studied the tidal straining effect on density stratification during the same 25.6-hr period along the same transect (Li et al., 2009). In this study, we conduct a numerical model simulation and illustrate that the lateral variations in the salinity and velocity fields are comparable or even larger than the vertical variations within a diurnal tidal cycle.

The density distribution within any estuary is a result of both advective and mixing processes. In Barataria Pass, the turbulent mixing is closely related to the magnitude of ebb/flood current and the strength of the tidal bottom boundary layer. Characteristics of horizontal advection processes in the inlet is that maximum flood currents are located at the central part of the deep channel for a large part of the flood period. This differential advection (Nunes and Simpson, 1985), when acting upon the along-channel density gradient, produces a distinct
density difference between the shoal and channel waters. In addition, the advection of Mississippi River water to the eastern channel during part of the flood period further enhances the density difference. On the contrary, maximum ebb currents swing between the western slope and central surface of the channel during the ebb. When maximum ebb flows are at the western slope, the differential advection mechanism does not work. When they go back to the channel center, the salinity contour lines are mostly horizontal due to weak vertical turbulence mixing. Thus, both situations are not favorable to produce an extreme density near the middle of the channel.

During flood period, when density distribution is high near the channel center and low at both shoals, the horizontal pressure gradient drives a lateral circulation with two counter-rotating cells and surface or near surface convergence. This result from the Barataria Pass is similar to that reported by Nunes and Simpson (1985). However, detailed analysis of momentum equations indicates that, in addition to the pressure gradient and vertical stress divergence, nonlinear advection and horizontal stress divergence are also important terms.

During ebb period, the lateral circulation is mostly eastward for the whole water column and persisting for almost the whole period. The surface divergence suggested by the differential advection mechanism is either non-existent or lasting for very short period. The main momentum balance across most of the transect is between the along-channel advection of cross-channel momentum and pressure gradient. In addition, the sectional averaged lateral velocity magnitude during ebb is comparable to that during flood, which is different from the idealized numerical experiment (Lerczak and Geyer, 2004).
Interactions among lateral circulation, along-channel tidal currents, and density stratification are complex processes. Lateral advection of momentum can act as an additional driving force for the estuarine circulation (Geyer et al., 2000; Lerczak and Geyer, 2004). The idealized numerical experiment with constant eddy coefficients (Lerczak and Geyer, 2004) demonstrated that the lateral circulation can be significantly different over a spring-neap cycle and density stratification can sometimes inhibit lateral circulation. Our study can be regarded as a starting point for further investigations of interactions among lateral circulation, estuarine circulation, and estuarine stratification in this partially stratified tidal inlet.
CHAPTER 4. TIDAL AND WIND-DRIVEN ESTUARINE-SHELF EXCHANGE THROUGH A TIDAL INLET: IMPLICATIONS FOR LARVAL RECRUITMENT

4.1. Introduction

Shelf-estuarine exchange is primarily controlled by tidal flows, large-scale inner shelf circulation, and small-scale local circulation near the mouths of estuaries or small tidal inlets. Understanding the detailed physical exchange processes and mechanisms across the continental shelf and coastal estuaries/bays interface is imperative in predicting the fate of pollutants and sediments (Yao et al., 2016), as well as the ingress of larvae of many marine species from shelf spawning areas into estuarine nurseries (Shaw et al., 1985b).

Tidal and density stratification are primary factors controlling the circulation and water properties in an estuary-inlet-shelf system on intratidal time scale. On the other hand, buoyancy force (e.g., river freshwater discharge) and winds are the predominant factors of subtidal circulation. Traditionally, lateral circulation in estuaries or tidal inlets is considered to be much weaker than the tidal currents. Recent observational and numerical modeling results reveal that these lateral motions can play an important role in estuarine-shelf exchange (Chen et al., 2009; Li and Li, 2012; Wu et al., 2018). Convergence (divergence) during flood (ebb) tides can generate intense vertical mixing (Wheless and Valle-Levinson, 1996), and affect particle transport and fish larvae distribution (Govoni et al., 1989).

River plumes are prominent features near river mouths due to buoyant water spreading over coastal waters. River plumes are widely reported and studied topic (Chao and Boicourt, 1986; Fong and Geyer, 2001, 2002; Garvine, 1974, 1977; O'Donnell, 1997; O’Donnell et al., 1998). Garvine (1974, 1977) presented observations of Connecticut River plume’s motion, which
showed strong horizontal shear at the front and strong density-driven surface flow normal to the tidal flow. Observations at the same area by O’Donnell et al. (1998) revealed a horizontal convergence at the plume front and significant downwelling. The structure, transport, and variability of river plumes are often significantly affected by wind stress and ambient current (Chao, 1988b; Fong and Geyer, 2001, 2002; Hetland, 2005; Lai et al., 2016). Generally, cross-shelf winds affect nearshore Ekman drifts by the sea level setup or setdown. Downwelling winds may cause the shoreward compression of the plume, while, upwelling winds cause seaward excursion of the plume (Chao, 1988b).

This study is motivated by idealized numerical experiments (Chao, 1988b; Kapolnai et al., 1996; Wheless and Valle-Levinson, 1996) which investigate the circulation and transport pathways associated with estuarine plumes under the influence of wind- and tidal-driven motions. Chao’s study focused on wide estuaries in which the Coriolis force is important, while the latter two studies considered a relatively narrow inlet (small Kelvin number), where rotational effects are unimportant. We will show that some results obtained from both wide estuaries and narrow inlets are also presented near the Barataria Pass, the area of focus of this study. Observational studies in other Northern Gulf of Mexico estuaries inspired the formulation of research questions in this study as well. Huguenard et al. (2016) conducted a field campaign in the Choctawhatchee Bay plume region near Dustin Inlet, Florida. They found that frontal processes dominated in overall mixing of the plume. Li et al. (2017) studied estuarine plumes using both in-situ measurements and synthetic-aperture radar imagery. In both studies, plumes showed asymmetry with western intensification. Huguenard et al. (2016) interpreted the asymmetry as a result of convergence caused by eastward ambient currents. By analyzing the momentum/vorticity equations, Li et al. (2017) attributed the Coriolis force as a dominate factor. We are curious to
identify the specific controlling mechanism in the Barataria Estuary plume. This paper is organized as follows. In section 4.2, we introduce study area and model configurations, and validate our numerical model. In section 4.3, we present shelf circulation response to wind events, evolution of the estuarine plume, and wind-driven variations of the plume. Lagrangian particle transport pathways are discussed in section 4.4. In section 4.5, several mixing mechanisms are proposed. Finally, in section 4.6, we summarize our results.

4.2. Material and methods

4.2.1. Study area

The Barataria Estuary is located in southeastern Louisiana. It is shaped with a major axis of ~ 110 km in the north-south direction and a width of ~ 50 km at its mouth (Figure 4.1). It is connected to the Louisiana Continental Shelf through several tidal inlets and is also connected to the Gulf Intracoastal Waterway in the upper estuary. The primary exchange (~66%) between the Barataria Estuary and the coastal ocean is through a tidal inlet, Barataria Pass (Marmer, 1948), which is 800 m wide. The mean water depth in the inlet is ~ 20 m with shoals 2-3 m deep on both sides. A ~ 50 m deep depression is located very close to the inlet inside the estuary. Similar to other estuarine bays in the Northern Gulf of Mexico, Barataria Bay is shallow, with an average depth of ~ 2 m. The water depth over the continental shelf varies, from ~ 4 m near the barrier islands, to ~ 10 m about 10 km offshore. The main freshwater source is the Davis Pond Diversion which was is built to divert Mississippi River water into the estuary to combat saltwater intrusion. Its maximum discharge capacity is ~300 m$^3$ s$^{-1}$. Tidal forcing is diurnal with a tropic tidal range of ~ 0.35 m at the mouth (Das et al., 2012).
4.2.2. Numerical model configuration

The Finite Volume Coastal Ocean Model (FVCOM, version 2.6) was used for this study (Chen et al., 2011). FVCOM is a three-dimensional, hydrostatic, free-surface, primitive-equation ocean model. The triangular grid in the horizontal can resolve complex bathymetric geometries. The model has been successful applied in a number of coastal (Huang et al., 2011), river plume (Lai et al., 2016), and estuarine studies (Xue et al., 2009). The computational domain covered from east of Mobile Bay, Alabama to west of Galveston Bay, Texas and offshore to about 27 °N. The horizontal resolution varied from 10 m in the upper estuary, to about 8 km near the open boundary with 19 vertically uniform sigma layers, which is ~ 0.1 m over the shoal and ~ 1m in the deep channel of the tidal inlet. The same model configuration was used for our spring 2010 and summer 2008 studies documented in the first two manuscripts. For the current study, a three-month simulation, from October to December 2007, was completed first. After that, a restart file was used to continue another three-month simulation, from January to March 2008. In this study, we mainly focused on the first 15 days of January, because there were two distinctly different wind events during this period. One wind event was N-NE wind while another was SE-S wind (Fig. 4.3).
Figure 4.1. (a) Mesh depicts the FVCOM model domain. (b) Model domain of the Barataria Esutury with bathymetry shown. The numbers 1-4 represent four passes: Caminada Pass, Barataria Pass, Pass Abel, and Ouatre Bayou Pass, respectively. Number 5 indicates Southwest Pass.
Figure 4.2. (a) Time series of daily mean discharge for the Mississippi River at Tarbert Landing, MS (black line) and 15-min interval discharge for the Davis Pond Diversion (red line) in 2008. (b) Identical to (a) but for only the first 15 days of 2008. Low discharge is 0-10,000 m³ s⁻¹, moderate discharge is 10,000-20,000 m³ s⁻¹, and high discharge is above 20,000 m³ s⁻¹.

Daily mean river discharge was obtained from the Mississippi River at Tarbert Landing station (USGS 07295100), which was linearly interpolated into 15-min intervals for consistency with other freshwater discharge station data collected at Atchafalaya River at Morgan City (USGS 07381600) and Wax Lake Outlet at Calumet, LA (USGS 07381590). The main freshwater source in the Barataria Bay is only a small fraction of the Mississippi River discharge diverted through the Davis Pond freshwater diversion. The discharge for the whole 2008 year and the first 15 days are shown in Fig. 4.2a and 4.2b respectively. Generally, the Mississippi River discharge exhibited a typical annual cycle, with highest flow in spring and lowest flow in late fall and early winter. Based on Walker et al. (2005) classification, discharge below 10,000
m$^3$s$^{-1}$ is considered to be low, discharge between 10,000 m$^3$s$^{-1}$ and 20,000 m$^3$s$^{-1}$ is moderate, and discharge above 20,000 m$^3$s$^{-1}$ is high. For our study period, the Mississippi River discharge was moderate, with a high of ~16,000 m$^3$s$^{-1}$ and low of ~11,000 m$^3$s$^{-1}$ (Fig. 4.2a). The first 12 hours was linearly interpolated from 0 to daily averaged 16,000 m$^3$s$^{-1}$, this may affect the development of the Southwest Pass plume. Thus, our analysis does not include the first 12 hours.

The Davis Pond Diversion discharge varied from 50 m$^3$s$^{-1}$ to 100 m$^3$s$^{-1}$ during the study period (Fig. 4.2b), with a sudden decrease to nearly zero on 10 January 2008.

The only atmospheric forcing included in the model was the 3-hourly 10-m wind, which was obtained from the NOAA National Centers for Environmental Prediction (NCEP) North American Regional Reanalysis (NARR) products and was interpolated into the entire computational domain. The comparison between NARR wind and National Buoy Data Center observed wind for SW Pass station is shown in Fig. 4.3. The station is located at the Mississippi River Southwest Pass while the Grand Isle station is located at the mouth of the Barataria Bay. The pattern of the NARR wind was consistent with observed wind. However, the NARR wind was underestimated at the Grand Isle station, which has been recently verified in Mariotti et al. (2018). The first 3 days in January were post-front period, with strong north and northeast winds. This pattern was followed by southeast winds, which lasted for about 6 days.

Model initial conditions were from the restart file of a three-month simulation run from 1 October 2007 to 31 December 2007. The open boundary conditions for sea level were derived from the NOAA tides and currents website (https://tidesandcurrents.noaa.gov/). Seal level time series of Dauphin Island, Southwest Pass, Freeport, and Galveston Pier 21 were directly used to prescribe sea level at the easternmost, southeastern, southwestern and the westernmost nodes,
respectively. Sea level elevations at other open boundary nodes were linearly interpolated from these four nodes.

The vertical eddy viscosity was calculated using the modified Mellor and Yamada level 2.5 turbulence model (Mellor and Yamada, 1982) and the horizontal eddy viscosity are given by Smagorinsky (1963) turbulence closure schemes. The time integration of the model used the split-mode time stepping method with a 2 s external time step and 0.2 s internal time step. The wet/dry scheme was turned on because the model included wetlands.

Figure 4.3. Three-hourly interval wind of buoy (b and d) and NARR (a and c) at Southwest Pass (a and b) and Grand Isle (c and d) for January, 2008.
4.2.3. Model validation

4.2.3.1. Water level elevation

Time series of modeled and observed water level elevations at six NOAA stations along the Northern Gulf of Mexico coast and six USGS stations in the Barataria Estuary are shown in Fig. 4.4 and Fig. 4.5, respectively. The time series cover three months from January 1 to March 31, 2008. Generally, the model results are in good agreement with the observations. The greatest discrepancy occurred at the NOAA Sabine Pass station, with the modeled magnitude 0.5 m greater than the measured one during flood tides. This may be attributed to the fact that the observation station is located behind an island, which was not resolved in the current model grid. In addition, the model not only predicted the tidal variation in the estuary well, but also the tidal decay in the upper-estuary propagation. For example, on 19 March 2008 (Fig. 4.5) the tidal signal decreased from stations in the lower estuary (Fig. 4.5b) to stations in the upper estuary (Fig. 4.5g).

The subtidal variations were also compared to demonstrate the reliability of the model. In order to guarantee the accuracy of filtering, a time series with no missing data was optimal. Thus, hourly water level elevation at USGS Grand Terre Island station and 3-hourly NARR wind at the nearest element near the Grand Isle were chosen. The time period was one year, however only first 15 days in January are presented here. The time series of signal over different frequencies (low frequency: \( T > 20 \) days; subtidal frequency: 40 hours < \( T < 20 \) days; tidal frequency: 4 hours < \( T < 40 \) hours, high frequency: \( T > 4 \) hours) are shown in Fig. 4.6. The discrepancy between modeled and observed variation for low frequency was less than 4 cm (Fig. 4.6a). The subtidal frequency of water level variations agreed well between the modeled and the observed.
The local wind induced setup and setdown were apparent (Fig. 4.6b). With northwest, north, and northeast winds, the subtidal water level elevation decreased (1-3 January, 2008), it increased with southeast and south winds (4-5 January, 2008). During the next 5 days, the wind varied from southeast and southwest and the water level elevation oscillated around maximum water level until the wind direction changed to coming from two northern quadrants. The modeled tidal phase agreed well with the observed one (Fig. 4.6c). The discrepancy in tidal amplitude was ~ 5 cm. The discrepancy for high frequency was as much as 10 cm (Fig. 4.6d).

Thus, the water level comparison indicated that the model was capable of reproducing the tidal and subtidal process in the Barataria Estuary, which was a necessary condition for the analysis of intra-tidal and inter-tidal variability of the estuarine plume.

For the 15-day study period, the wind observation at the National Data Buoy Center (NDBC) Grand Isle station only recorded the last 5 days. The comparison results show that they were consistent in both direction and magnitude (Fig. 4.6e).

4.2.3.2. Salinity

Three-month time series of hourly observed and modeled surface salinity at USGS Grand Terre Island are shown in Fig. 4.7. Generally, the modeled salinity showed similar trends with observed salinity, especially over the first 15 days, which was our time period of interest.
Figure 4.4. (a) Time series of wind vector at Grand Isle and (b-g) comparison of 6-min interval observed (black) and hourly modeled (red) water elevations at six coastal NOAA stations.
Figure 4.5. (a) Time series of wind vector at Grand Isle and (b-g) comparison of hourly observed (black) and modeled (red) water elevations at six USGS stations in Barataria Estuary. Shaded areas indicate missing values.
Figure 4.6. Band-pass filtered observed (black) and modeled (red) water level elevation at USGS Grand Terre Island station and 10-m wind at NDBC Grand Isle station (orange) and NARR (blue). (a) low frequency: $T > 20$ days, (b) subtidal frequency: 40 hours $< T < 20$ days, (c) tidal frequency: 4 hours $< T < 40$ hours, (d) high frequency: $T > 4$ hours, (e) raw data.

Figure 4.7. Time series of hourly observed (black) and modeled (red) surface salinity at USGS Grand Terre Island station for January, February, and March 2008.
4.3. Results

4.3.1. Features of modeled currents on the seaward of the inlet

Flows during flood tides were almost uniformly toward the inlet within 2 km from the inlet (Fig. 4.8a), similar to a potential flow (Stommel and Farmer, 1952). Further away, flows mostly came from the east of the inlet. During ebb tides, flows were also alike potential flow with a 2 km radius semi-circle near the inlet, and defected toward the right-hand side further offshore (Fig. 4.8b). There was asymmetry in the speed of maximum flood and ebb currents, with the latter about 30% greater than the former. This was obviously due to the discharge from the Davis Pond Diversion. For the idealized inlets (Chao, 1990; Wheless and Valle-Levinson, 1996), the most prominent feature in residual flow was two counter-rotating eddies located on either side of the ebb flow. In our case, these two eddies also occurred, but were much weaker (Fig. 4.8c). This may have been influenced by the variation of bathymetry along the inlet. In the idealized cases, the water depth in the inlet was constant and shallower than that over the shelf. For the Barataria Pass, the average water depth was ~ 20 m in the inlet with a depression up to ~ 50 m just north of it. This depression was greater than 4 m in depth and was next to the inlet over the continental shelf. As a result, the deep topography decelerated the incoming tidal current toward the inlet during flood tides. During ebb tides, the seaward currents entering the inlet also decelerated more due to increasing water depth. Thus, the velocity shear and vorticity were much weaker in the Barataria Pass.

4.3.2. Lateral variability in the inlet

In Chapter 3, we have demonstrated that significant lateral variability in salinity and secondary circulation within a tidal cycle in Barataria Pass during the summer. This study
demonstrated that this lateral variability persists in the winter season and the patterns are similar (Fig. 4.9 and 4.10). Saline shelf water encroached near bottom of the inlet during flood tide (Fig. 4.9a) due in large part to the maximum along-channel velocity, which was located at mid-depth (Fig. 4.9c). The lateral circulation showed convergent flow in the deep channel, which can be attributed to differential advection (Lerczak and Geyer, 2004; Nunes and Simpson, 1985). That is, the strongest along-channel velocity in the deep channel advected saline water faster in the deep channel than that in the shoals. As a result, a cross-channel pressure gradient developed that drove a lateral flow. Compared with the lateral circulation in the summer case, this lateral circulation during flood tide was also controlled by recirculated estuarine water previous transported seaward on the prior ebb tide. When the western water was fresher than the eastern side, the western anticyclonic circulation was stronger than the western cyclonic circulation, and vice versa. At flood slack, the lower water column of the deep channel became relatively well-mixed, developing a weak stratification near surface within 4 m depth with no lateral flow greater than 0.1 m s\(^{-1}\) (Fig. 4.9b). During ebb tide, fresher estuarine water begins to transport on the downstream side of the inlet after flood slack. The maximum ebb currents occurred on the western slope of the inlet (Fig. 4.10c). The water column was vertically well-mixed but with a weak lateral gradient. The vertical velocity field showed upwelling appearing on the western slope (Fig. 4.10a), where maximum outflow occurred (Fig. 4.10c). At ebb slack, the weak surface stratification redeveloped (Fig. 4.10b). During ebb tide, lateral circulation was much weaker than that during flood tide.
Figure 4.8. Vertically integrated velocities near the inlet at (a) max flood and (b) max ebb. Current vectors were selected in a search radius of ~ 200 m and velocity greater than 2 cm s$^{-1}$. (c) Vertically integrated residual currents (averaged over one month, January 2008). Current vectors were selected in a search radius of ~ 200 m and velocity greater than 0.5 cm s$^{-1}$. 
Figure 4.9. A transverse section of salinity and lateral circulation (a and b) and along-channel velocity (c and d) in the Barataria Pass for maximum flood (a and c) and flood slack (b and d). View is looking into the inlet from the ocean. Vectors show lateral circulation. Colors represent salinity and salinity contour lines are superimposed. Solid black lines represent along-channel velocity.
4.3.3. The response of surface circulation on the shelf to different wind events

Winds, freshwater discharge and tides are the major forcing factors that control the development of estuarine plumes (Chao, 1988a, 1988b, 1990). During our investigation time (1 to 15 January, 2008), the Davis Pond discharge varied around ~70 m$^3$ s$^{-1}$ (Fig 4.2b). Thus, we
consider the other two factors the major variables during time period. Before we illustrate the behavior of the estuarine plume, which will be given in the next section, the surface currents response to different wind events in the Louisiana Bight will first be described. This wind-driven circulation affected the evolution of the plume over the inner continental shelf.

From 1 to 15 January, the wind varied from north/northeast to southeast and later returned back to northeast (Fig. 4.3). The first three days showed a typical characteristic wind of post frontal passage, namely, strong north and northeast winds (> 10 m s\(^{-1}\)). Another cold front occurred around 11 January, 2008, which introduced a clockwise rotary wind field (Fig. 4.3).

North/Northeast winds

Figs. 4.11 and 4.12 show surface currents and salinity under north/northeast winds, which were at the beginning and end of the ebb period, respectively. The magnitude of wind was greater than 10 m s\(^{-1}\). The development of the estuarine plume during this ebb period will be discussed in the next section to illustrate the plume response to the seaward wind. The surface currents demonstrated a clockwise gyre in the Louisiana Bight, which was related to the Mississippi River plume and has been confirmed by previous observations (Walker et al., 2005). The outflow from the Southwest Pass and South Pass developed a strong (40-60 cm s\(^{-1}\)) westward flow, Louisiana Coastal Current. Westward flow turned northwestward toward the coast near \(x=800\) km in Fig. 4.11, and separated into western and eastern branches around 3200 km in the \(y\) direction. The clockwise gyre brought outflow from the Southwest Pass and estuarine water along the shore back to the west of the delta. The coast currents immediate offshore of the barrier islands had two branches. West of the Barataria Pass, the current flowed along the coastline and later joined the westward Louisiana Coastal Current. East of the Barataria
Pass, the current flowed eastward and southeastward and joined the gyre in the Louisiana Bight. At the end of the ebb tide, the eastern branch of the gyre encountered the estuarine plume from the Barataria Pass (Fig. 4.12). This flow may have limited the spreading of the estuarine plume. More of this theory details will be discussed in the following next section.

East Winds

When the wind switched from the northeast to the east, the clockwise gyre in the Louisiana Bight was not obvious (Fig. 4.13). The westward Louisiana Coastal Current retreated to within 10 km from the Southwest Pass but still curved to the northwest and separated into two branches. The eastward branch was weak. Near the west bank of the Birdfoot delta, there was a westward flow that was a direct result of the westward wind direction.

Southeast Winds

Due to the Ekman transport, the outflow from the Southwest Pass moved along the coast to the right-hand side (Fig. 4.14). This westward flow increased compared to that under the east wind. The westward current near the Southwest Pass existed, but it was narrower and weaker than during the northeast wind. The flow pattern is consistent with a previous study (Walker et al., 2005). The two branches of coast currents combined to develop westward currents.

South/Southwest Winds

During the south wind, the gyre was not obvious (Fig. 4.15) while during the southwest wind, the westward Louisiana Coastal Current totally disappeared (Fig. 4.16). The northwest flow moved shoreward and separated into two branches. The left one generated a cyclonic eddy. The Ekman transport increased the right-hand side currents. This eastward flow together with
westward coast currents trapped the saline water on the western side of the delta. The flow on the western side of the Barataria Pass moved eastward and encountered the westward flow from the eastern side near the outflow from Barataria Bay.

West Winds

A pair of counter-rotating eddies still existed (Fig. 4.17). Surface waters along the Northern Gulf of Mexico coastline moved eastward and offshore toward the Southwest Pass in the Louisiana Bight.

Northwest/North Winds

The northwest/north winds (Fig. 4.18, Fig. 4.19) efficiently forced surface waters offshore toward the Southwest Pass. When encountering the outflow from the Southwest Pass, the flow curved westward and toward northwest. This northwestward flow turned southwestward when meeting the coast currents. The flow developed an “S” shape.

The characteristics of shelf circulation during different wind events were demonstrated in a study by Walker et al. (2005). Their findings were very similar to our simulated results. This indicates that shelf circulation is repeatable under similar environmental conditions. Murray (1972) showed that surface currents are more related to the magnitude of the wind and time history. Our model also reproduced this feature. For example, when the magnitude of northeast winds decreased, the clockwise gyre became weaker (figure not shown).
Figure 4.11. Surface currents and salinity at 03:00 January 3, 2008, at the beginning of the ebb tide. The white arrow indicates northeast wind.
Figure 4.12. Surface currents and salinity at 15:00 January 3, 2008, near the slack of the ebb tide. The white arrow indicates northeast wind.
Figure 4.13. Surface currents and salinity at 06:00 January 4, 2008. The white arrow indicates east wind.
Figure 4.14. Surface currents and salinity at 00:00 January 5, 2008. The white arrow indicates southeast wind.
Figure 4.15. Surface currents and salinity at 15:00 January 10, 2008. The white arrow indicates south wind.
Figure 4.16. Surface currents and salinity at 00:00 January 11, 2008. The white arrow indicates southwest wind.
Figure 4.17. Surface currents and salinity at 00:00 January 5, 2008. The white arrow indicates west wind.
Figure 4.18. Surface currents and salinity at 09:00 January 11, 2008. The white arrow indicates northwest wind.
Figure 4.19. Surface currents and salinity at 12:00 January 11, 2008. The white arrow indicates north wind.

4.3.4. The evolution of the estuarine plume within a diurnal tidal cycle

Following Garvine (1995), the Kelvin number ($K$) is calculated to determine the importance of the rotation effect, which is the ratio of the plume width (i.e., the width of Barataria Pass, L) to the internal Rossby radius $R_0$ ($R_0 = \sqrt{g' h / f}$). Here $g'$ is the reduced gravity acceleration, $h$ is the plume depth, and $f$ is the Coriolis parameter, taken to be $7 \times 10^{-5}$ s$^{-1}$ here. In our model, the plume width was about 1 km, the plume depth was ~ 4 m. Considering
a density difference of 4.5 kg m\(^{-3}\), the Rossby radius was \(\sim 6\) km. The Kelvin number was \(O(0.16)\) or less, which means the deflection of the outflow was due to alongshore transport rather than rotational effects.

Here we describe the evolution of the estuarine plume within a tidal cycle starting at the beginning of the ebb tide under north/northeast winds. At this time the wind was from north and, thus should be favorable to the spreading of plumes over the continental shelf. Fig. 4.20 through Fig. 4.25 show plume development over time. At the beginning of ebb tide, the freshwater was retained on the western shore by flood tide and thus, the western side of the inlet was fresher than the eastern side aligning the channel on the shore side (Fig. 4.20a). Along the transect salinity showed weak stratification at depth and upper water moved seaward, while the bottom water featured a weak landward flow (Fig. 4.20b). As the tidal cycle advanced toward maximum ebb stage, the buoyant estuarine water was advected seaward for a distance \(\sim 6\) km. Within 3 km from the inlet, flow was unidirectional and seaward at all depths. The water column was vertically well-mixed. From 3 km to 6 km, however, flows were separated at 4 m in depth, with the upper freshwater layer seaward, the bottom saline water landward (Fig. 4.21b), and a bottom–surface salinity difference up to 6. The surface buoyant fluid exited the inlet as a potential flow and spread radially to form an expanding plume (Fig. 4.21a). The alongshore plume length was \(\sim 11\) km. Note that there are three others tidal inlets along the coast, each with estuarine plumes interacting with one another. Therefore, the dynamics of plume in the estuary are even more complex. But for this study, we only focused on the plume from the main pass, Barataria Pass. At the late ebb period, the outflow in the plume turned toward the west (Fig. 4.22a). Since rotation effect is unimportant, this deflection was driven by a westward alongshore current caused by the northeast wind (Fig. 4.12). Similar to the idealized estuary (Wheless and
Valle-Levinson, 1996), after exiting the inlet, this outflow lifted off the bottom and remains detached from the shelf all the times (Fig. 4.21b, Fig. 4.22b, and Fig. 4.23b). When the tides approached slack before flood, the tidal velocity decreased greatly (Fig. 4.23a). However, there was still substantial flow due to the outflowing gravity currents. The plume extended downstream. This asymmetry, namely the right-hand side spreading stronger than the left-hand side when looking seaward, was caused by wind forcing. Details considering relaxation of winds will be discussed in the next section.

As tidal forcing reversed to maximum flood tide, the landward flow transports shelf saline water toward the inlet. The maximum flood current occurred near the upstream side of the inlet (Fig. 4.24a) and at the surface (Fig. 4.24b). Because of the tidal mixing and the landward advection of the saline shelf water, the inlet tended to be well-mixed and weakly stratified. Note that in chapter 3 we discussed the stratification in the inlet, located ~2 km away from the head of the transect indicated by magenta color, where the water column was stratified (well-mixed) during flood (ebb) tide. The mechanism was dominated by the nonlinear advection. However, for this shallow continental shelf, the driven forcing may be due to the tidal straining, which caused stratification (well-mixed) during ebb (flood) tide. A substantial area of surface buoyant water and a sharp salinity front with strong surface current remained on the downstream area. Shelf flow became weak, this may be related to wind-driven circulation.
Figure 4.20. (a) Horizontal distribution of surface salinity and velocity vectors at the beginning of ebb tide (3:00 January 3, 2008) under northwest wind. The contour interval for salinity is 1 and the outmost contour line is 29. The blue lines are 10-m and 20-m isobaths, respectively. The vertical integrated along-channel velocity in the Barataria Pass is shown in the upper-left panel and the red dot represents the time instance at which the figure is shown. (b) Cross-sectional distribution of salinity and circulation along the transect indicated by magenta color line in (a).
Figure 4.21. Same as Fig. 4.20 but at 11:00 January 3, 2008 (GMT), maximum ebb tide.
Figure 4.22. Same as Fig. 4.20 but at 13:00 January 3, 2008 (GMT), near ebb slack.
Figure 4.23. Same as Fig. 4.20 but at 16:00 January 3, 2008 (GMT), the beginning of flood tide.
Figure 4.24. Same as Fig. 4.20 but at 21:00 January 3, 2008 (GMT), near maximum flood tide.
Figure 4.25. Same as Fig. 4.20 but at 05:00 January 4, 2008 (GMT), near flood slack.
4.3.5. Wind-driven variations of the estuarine plume

In an idealized study, plume behavior is only driven by tides and buoyant discharge (Kapolnai et al., 1996; Wheless and Valle-Levinson, 1996) and the time evolution of the outflow and frontogenesis is repeatable during each tidal cycle. Significant subtidal circulation existed in the Louisiana Bight which was mainly driven by winds. If the upstream Davis Pond freshwater discharge rate is almost constant, it will surely add an important subtidal component to the quasi-periodic tidal variation of the estuarine plume. The wind-driven variations of an estuarine plume in a wide estuary (earth’s rotational effect is important) was numerically studied by Chao (1988b), using a three-dimensional, primitive-equation and rigid-lid model. His model results are described here. The upwelling-favorable wind caused the seaward excursion of the plume and drastically weakened the stratification. The density current opposed the wind, which made the coastal jet unlikely to develop. The downwelling-favorable wind induced a wind-driven coastal jet inside the light water pool, which was on the downstream side when looking seaward in the north hemisphere. For the cross-shelf winds, the seaward wind increased the freshwater export onto the shelf, enhanced the stratification, and reduced the Ekman drift nearshore. On the other hand, the landward wind induced the withdrawal of freshwater from the shelf, weakened the nearshore stratification, and therefore enhanced the Ekman drift. His study focused on wide estuaries, in which the Coriolis force plays an important role. In our case, we have verified that rotational effect is unimportant in the plume. However, we have also shown that shelf circulation is primarily wind-driven. This circulation can be considered as ambient current (Fong and Geyer, 2002) and also contributes to the development of estuarine plume.

In order to examine how wind affects the estuarine plume, another simulation, in which surface wind was turned off, was conducted from 1 January to 31 March, 2008. Fig. 4.26 through
Fig. 4.29 show features of surface plumes with and without wind forcing. The first impression is that the Coriolis force was inconsequential for the plume, since the downwind coastal jet did not exist for downwelling-favorable wind, which is a significant feature in the wide estuaries (Chao, 1988b). The deflection of the plume was mainly caused by the local wind. Without winds, the plume was symmetric relative to the axis of the Barataria Pass.

The responses forced by cross-shelf winds were as expected. Fig. 4.26 shows surface features under seaward wind. More freshwater was exported onto the shelf in this simulation. The 29 isohaline was forced across the 10-m isobath. The cross-shore length of the plume was about 14 km. The flow at the seaward edge of the plume turned sharply anticyclonically, which was also shown in the idealized case (Kapolnai et al., 1996). The along-shore component of the wind made the plume not symmetric with respect to the inlet mouth: the spreading was larger on the western side compared with that on the eastern side. Chao (1988b) pointed out that seaward wind enhances the freshwater export from the estuary but is ineffective in advecting the plume downstream. This was also true in our model. Relaxation of the wind caused a decrease in the freshwater export, which can be seen from the salinity difference near the inlet. The water was fresher with seaward wind (Fig. 4.26a). The plume structure of wind relaxation was not significantly different from the realistic case. By contrast, the landward wind suppressed the outflow of freshwater. The wind-induced landward currents transport shelf water toward the inlet, resulting in the 29 isohaline moving shoreward of the 10-m isobath, about 5 km from the inlet mouth (Fig. 4.27a). The wind relaxation lead to the expansion of the plume. The shoreward length was about 10 km (Fig. 4.27b).

As stated previously, the Coriolis force was insignificant in the plume. Theoretically, downwelling (upwelling) favorable winds can only advect the plume downstream (upstream).
However, the 3-D momentum balance indicated that the Coriolis force had the same order of magnitude with pressure gradient over the shelf. As a result of Ekman transport, downwelling-favorable winds (northeast/east winds) forced surface waters shoreward (Fig. 4.12 and Fig. 4.13), resulting in the plume retreating toward the inlet. Upwelling-favorable winds (southwest/west winds) forced surface waters eastward and offshore (Fig. 4.16 and Fig. 4.17) along the Northern Gulf of Mexico coastline, resulting in spreading of the plume. Fig. 4.28a shows plume structure during the downwelling-favorable wind. The 28 isohaline almost intruded across the 10-m isobath. More freshwater was retained inside the estuary. The downstream component of wind made the western side of the plume larger than the eastern side. Wind relaxation enhanced the freshwater export from the estuary (Fig. 4.28b) and makes the plume symmetric. The anticyclonic circulation at the plume edge switched to cyclonic circulation (Fig. 4.29a) under the upwelling-favorable winds. The plume was wider on the eastern side than on the western side. When turning off the wind, the anticyclonic circulation recovered and the plume was more or less symmetric (Fig. 4.29b).
Figure 4.26. Surface salinity and currents under (a) seaward wind and (b) no wind. The vertical integrated along-channel velocity in the Barataria Pass is indicated in the upper-left panel. The red dot represents the time instance at which the figure is shown. The magnitude of wind is 8.3 m s\(^{-1}\). The white contours are salinity, the interval is 1, and the contour lines to the south represent 29 in (a) and 28 in (b). The blue contour lines are 10-m and 20-m isobaths, respectively.
Figure 4.27. Same as Fig. 4.26 but under (a) landward wind and (b) no wind. The magnitude of wind is $4.6 \text{ m s}^{-1}$. 
Figure 4.28. Same as Fig. 4.26 but under (a) downwelling-favorable wind and (b) no wind. The magnitude of wind is 7.3 m s\(^{-1}\).
Figure 4.29. Same as Fig. 4.26 but under (a) upwelling-favorable wind and (b) no wind. The magnitude of wind is 6.6 m s$^{-1}$. 
4.4. Bay-shelf exchanges

The characteristics of transport through the inlet and bay-shelf exchange process are studied by Lagrangian Particle Tracking. A group of 2,880 particles, Group A, were seeded just offshore of the inlet within 10-m isobaths. Another group of 2,400 particles, Group B, were placed over the shelf between 20-m and 60-m isobaths (Fig. 4.30). Initially, the particles were located at the surface, mid-depth, and near bottom, and released at the beginning of flood tide under two different types of wind events, northeast (seaward, denoted as T1) and southeast (landward, denoted as T2) winds. Only horizontal velocity components of the flow were used in computing the particles’ trajectories. Velocities at the same sigma level were used for the entire time period. To examine how exchange processes, respond to wind-driven circulation in the vicinity of the inlet, additional particle tracking simulations, with current field from zero surface wind simulations were also performed for the same period.

4.4.1. Transport through the tidal inlet within a diurnal tidal cycle

For seaward winds (i.e., T1), the maximum tidal excursion during the flood period for passive particles placed at the surface was approximately 6 km (Fig. 4.31a). Almost all particles that entered the bay were transported northeastward. The maximum flood excursion for particles placed at the mid-depth was comparative to that at the surface (Fig. 4.32a). The maximum flood excursion for particles placed near bottom was slightly smaller, ~4 km (Fig. 4.33a), presumably due to reduced tidal currents near the bottom. For no wind simulations, the maximum flood excursion had little change (Fig.4.31b, Fig.4.32b, Fig. 4.33b) compared to the seaward wind cases. However, percentage of particles entering the Barataria Bay was about 10% more than that for the seaward winds. All the particles that entered the inlet during the flood tide were expelled.
offshore during the next ebb tide. Particles at the surface had a maximum offshore displacement of ~ 10 km toward the plume edge (Fig. 4.31c, red dots). Particles at mid-depth and near bottom were retained within 6 km and 4 km distance from the barrier islands, respectively (Fig. 4.32c, Fig. 4.33c, red dots).

Figure 4.30. Initial location of particles for nearshore (group A, red) and shelf (group B, blue). The tidal phases at T1 and T2 are indicated in the upper-left panel.

For landward winds, almost 50% more particles at the surface could enter the bay during the flood tide compared with that for the seaward winds (Fig. 4.31a, T2). The maximum bayward excursion was about 12 km. As a result of wind-driven circulation, some of the particles could be transported northwest. Particles remaining near seaward of the inlet were pushed in the downstream direction of Kevin wave propagation. Similar patterns were also present for particles staying at mid-depth and near the bottom (Figs. 4.32a, 4.33a, red dots). Without wind forcing, the percentage of particles entering the bay decreased by about 20% for particles at the surface.
and near bottom, but the percentage of mid-depth has no noticeable variation (Figs. 4.31b, 4.32b, 4.33b, blue dots). During the next ebb tide, some particles that entered the bay were expelled out. However, a large portion of particles were retained in the bay, especially at the surface and at mid-depth (Fig. 4.31c, Fig. 4.32c).

In summary, for both seaward and landward winds, particles initially at mid-depth entered the Barataria Bay more easily than those placed at the surface and near the bottom. This was also true for our no wind experiment. Landward winds were favorable for particle ingress, increasing by at least 40% at the mid-depth. For seaward winds, particles struggled to settle in the bay. The majority of these particles were expelled offshore in the next ebb tide, no matter which depth they were located. This result is consistent with other studies (Brown et al., 2000; Kapolnai et al., 1996). Particles at the mid-depth and near bottom were effectively retained in the estuary (Fig. 4.32d, Fig. 4.33d).
Figure 4.31. Particle locations, for nearshore group (Fig. 4.30 red dots) placed at the surface and released at time T1 (red) and T2 (blue) (Fig. 4.30 inserted panel), at the end of initial flood tide (a) with wind and (b) without wind, and at the end of initial ebb tide (c) with wind and (d) without wind.
Figure 4.32. Same as Fig. 4.31 but for particles placed at the mid-depth.
Figure 4.33. Same as Fig. 4.31 but for particles placed at the bottom.
4.4.2. Preliminary simulations on shelf-estuarine exchange

Simulations were performed to determine shelf-estuarine exchange. Passive particles were placed at the surface between 20-m and 60-m isobaths near the Mississippi River Delta (Fig. 4.30, blue dots), which is the potential spawning area for Gulf menhaden (Shaw et al., 1985a). Particles were released during northeast (T1) and southeast winds (T2) then tracked for 30 days. T1 was 45 hours ahead of T2, thus, these simulations have a long overlapping period. Based on the animation, initially, particles were controlled by Louisiana Bight circulation. However, once they entered the westward Louisiana Coastal Currents, they were transported westward toward the Terrebonne-Timbalier Bay and Texas shelf (Fig. 4.34). This confirms the assumption of Shaw et al. (1985b) that Gulf menhaden larvae would be carried west-northwest along western Louisiana. They also assumed that, once near shore, larvae could be transported into the estuaries by tide-driven and wind-driven circulation or other shelf-estuary exchange mechanisms. These claims are not evident in our simulation. Further discussion on this topic is provided in the next section.

Figure 4.34. Particle locations, for shelf group (Fig. 4.30 blue dots) placed at the surface and released at time T1 (red) and T2 (blue) (Fig. 4.30 inserted panel), after 30 days.
4.5. Discussion

4.5.1. Mixing in the inlet

Mixing in the inlet can be caused by unstable stratification and vertical shear of velocity. During ebb tide, lighter estuarine fluid is advected over the more saline fluid so that the water column becomes stratified. Flood tidal currents advect saline water into the inlet with surface water moving faster than the bottom water, resulting in unstable stratification and vertical mixing. This phenomenon is called tidal straining (Simpson et al., 1990). However, we have confirmed that this situation does not exist in our numerical simulations. In the Barataria Pass, the water column became well-mixed during ebb tide while stratified during flood tide. The dominant mechanism is non-linear advections. Mixing was much stronger during ebb tide than during flood tide, with maximum vertical eddy viscosity ~ $O(10^{-1})$ m$^2$ s$^{-1}$, which is an order of magnitude larger than that during flood tide.

Following Wheless and Valle-Levinson (1996), the gradient Richardson number ($R_i = \frac{-g \frac{\partial \rho}{\partial z}}{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2}$) is used to measure the sensitivity of the flow to mixing, due to the growth of interfacial instabilities (i.e., turbulence). This nondimensional parameter represents the ratio of vertical stratification to velocity shear at a particular depth. The critical $R_i$ value is 0.25. If $R_i$ is below this value, interfacial instabilities can grow and cause mixing. At a point located in the middle of the inlet $R_i$ was calculated at maximum flood (Fig. 4.9a) and maximum ebb (Fig. 4.10a). The results show that small $R_i$ was confined to within 5 m from the bottom during flood tide (Fig. 4.35a), resulting in a well-mixed layer near the bottom and stratification on the top (Fig. 4.9a). During ebb tide, the small $R_i$ occurred in the lower half of the water column, (i.e. below 10
m in a 20 m deep water column) (Fig. 4.35b), resulting in an almost vertically well-mixed water column (Fig. 4.10a).

![Graph showing depth vs. Richardson number](image)

Figure 4.35. Gradient Richardson number in the FVCOM simulation at a point in the center of the Barataria Pass at maximum flood and maximum ebb tides. Dashed line shows where $R_i = 0.25$. If $R_i < 0.25$, it indicates that the area is susceptible to enhanced mixing due to shear instability.

4.5.2. Mixing in the estuarine plume front

Small-scale fronts associated with the mixing of brackish water and seawater can create regions of strong convergent surface flows and downwelling. Therefore, these fronts are of great importance to the shelf–estuarine exchange by affecting transport and dispersion of surface material near the plume front. Huguenard et al. (2016) conducted a field campaign to assess mixing processes of the estuarine plume in the Choctawhatchee Bay, located in the northeast Gulf of Mexico. They found that frontal processes dominated (59%) in overall mixing. On the western Louisiana continental shelf, Li et al. (2017) also observed estuarine plume outside of
constricted channels of the Sabine and Calcasieu Lakes. They found that the plume formed during multiple tidal cycles, and sustained during the flood tides as a result of weak diurnal tides.

Garvine (1984) developed a time dependent model applicable to a radially spreading plume. The model results suggest that there are two specific sites of mixing in the plume: one at the leading surface front and the other at an internal hydraulic jump or bore that forms due to interfacial wave coalescence. Froude number, \( F_r = \frac{u_1}{\sqrt{g' h_4}} \) (variables are shown in Fig. 4.36), deals with the relationship between gravity and inertial forces. For \( F_r < 1 \) the flow is considered subcritical flow; while for \( F_r > 1 \), the flow is characterized as supercritical flow. Our model also generated interfacial instability, which was predominantly concentrated at the sloping frontal interfacial region just behind the frontal convergence (Fig. 4.22b). Considering \( g' = 0.065 \text{ m s}^{-2}, h_4 = 4 \text{ m}, \) and \( u_1 = 0.2 \text{ m s}^{-1} \), we obtained \( F_r = 0.39 \). Therefore, the flow inside the gravity head was subcritical, which would allow an upstream return flow and allow interfacial waves to coalesce some distance behind the front. This internal jump also existed in the experiment of Rottman and Simpson (1983) and in the field observations (Pritchard and Huntley, 2002). These two different experiments (Rottman and Simpson, 1983; Garvine, 1984) showed similar internal structures. Pritchard and Huntley (2002) postulated that this downstream hydraulic jump is independent of the upstream flow conditions and not affected by modulations in the source flow from the estuary.
Figure 4.36. Schematic diagram illustrating the mixing at the head of a gravity current (Pritchard and Huntley 2002), where $u_1$ is the frontal velocity, $u_4$ is the buoyant flow overtaking (relative to the front) velocity, $h_1$ is the total water column depth, $h_4$ is the thickness of the buoyant layer, $\rho$ is the density of seawater, and $\Delta \rho$ is the density anomaly between plume and ambient seawater.

4.5.3. Coastal downwelling/upwelling

Fig. 4.37 and Fig. 4.38 show local upwellings and downwellings at the surface layer, parallel to the isobaths, which could be one of the mechanisms accounting for cross-shelf exchange (Ladd et al., 2005). The model results show that this phenomenon is irrelevant to wind stress because also occurred in the simulation without wind forcing. There are two potential mechanisms that could generate it: (1) Mississippi River plume generated internal waves; (2) interaction between Mississippi Canyon flows with shelf circulation (Chen and Allen, 1996; Lafuente et al., 1999). The dynamics of these upwellings/downwellings are not the topic of this study, however this phenomenon demonstrates the need of further studies.
Figure 4.37. Distributions of vertical velocity at the surface and at t=41 (hours) starting from simulation.

Figure 4.38. Distribution of vertical velocity at the surface and at t=61 (hours) starting from simulation.
4.6. Implication for fish larvae recruitment

Many ecologically important fish species spawn eggs offshore or over the inner continental shelf during fall/winter season. Their larvae are then transported into coastal and estuarine nursery grounds. Cushing (1975) found that the abundance of the fishery stock is highly related to the processes by which fish larvae are transported into estuarine nursery grounds. There are two major phases of movement necessary for the recruitment in estuarine nursery grounds (Boehlert and Mundy, 1988). The first phase is accumulation of larvae in the nearshore zone, while the other is accumulation of larvae near inlets and estuary mouths. Therefore, recruitment would not be possible without the help of favorable currents. Based on data collected during three consecutive winters, Shaw et al. (1985b) hypothesized that Gulf menhaden larvae are passively transported west-northwest along western Louisiana by the mean circulation. If favorable astronomically and meteorologically driven flows are paired with a patch of larvae near inlets or estuarine mouths, transportation into the estuaries is possible. They also found that larvae do not necessarily recruit to estuaries nearest to their offshore spawning areas.

The net flow of Barataria Estuary is seaward because of the Davis Pond Diversion. This creates the risk of larvae transport out of the estuary and away from their nursery areas. The results from our numerical simulations indicate that several mechanisms may bring particles to mid-depth and below: (1) secondary circulation in the tidal inlet; (2) flow convergence and downwelling in the front of estuarine plumes; (3) downwelling along the coast. Once at depth, these particles could be advected by landward currents in the lower layer.
4.7. Summary

The FVCOM model was run with realistic bathymetry, discharge, local and remote wind forcing, stratification and tides. The model-predicted water level elevations are comparable to observations at stations near the coast and in the Barataria Estuary. The comparison results indicate that the model can accurately reproduce tidal and subtidal constituents. The model predicted that offshore surface circulation is primarily controlled by local winds, consistent with previous studies (Oey, 1995; Walker et al., 2005). The model reproduced the anticyclonic gyre in the Louisiana Bight and showed that this clockwise circulation was modified by winds, with northeast winds intensifying it and northwest winds weakening it.

The modeled estuarine plume response to winds was analyzed under different wind events and compared to simulations without wind. The results indicate that the wind-driven coastal circulation determine the first-order plume response. The cross-shelf winds controlled the spreading of the plume in the along-channel direction, while along-shelf winds determined the along-shore development. Without winds, the plume was aligned symmetrically with the channel. There was an internal hydraulic jump on the plume interface, which resulted in instability and mixing. The offshore circulation also controlled the development of the plume, with downwelling-favorable wind inducing circulation retarding it, while up-welling favorable wind promoted seaward spreading.

We have demonstrated the performance of our FVCOM estuarine-shelf circulation model. Major large-scale circulation features seem well reproduced, although more model validation is desirable due to the sparsity of current available observations. Therefore, it would be too early to assert “agreement” with observations. In addition, there are many small-scale physical
phenomena, (i.e. mixing and dispersion of the plume frontal zone, estuarine plume-related internal wave generation, and mechanisms for coastal upwelling/downwelling) which deserve future investigations. We see several important areas for future model-assisted research:

(1) Coastal current, - including the dynamic interaction of local (i.e. tides, local wind and river discharge) and large-scale (i.e. Louisiana Bight gyre) physical features and their influence on the location and strength of the various branch points;

(2) Bay-Shelf exchange – including the relative roles of wind, baroclinicity, along-shelf transport, and Gulf Steam features in regulating these critical exchanges.

(3) Oceanic internal waves and their role in the interaction between large-scale tides and smaller-scale turbulence.
CHAPTER 5. SUMMARY AND CONCLUSIONS

This dissertation examined the impact of river diversions on the estuarine salinity gradient and residence time, investigated the mechanisms of lateral circulation in the tidal inlets, and studied characteristics of circulation, estuarine plumes, and particle transport through a narrow inlet and over the continental shelf with numerical model.

In chapter 2, estuarine salinity gradient and residence time were studied using a three-dimensional, high-resolution, hydrodynamic model under different discharge scenarios (the actual 2010 Davis Pond discharge, no discharge, and a constant \( \sim 850 \, \text{m}^3 \, \text{s}^{-1} \) discharge of the Mid-Barataria Diversion). The numerical model was validated using observed water level elevations and salinity distribution in space and time. In this study, the three-month average salinities indicated that surface salinities had less variation in the DPD scenario compared with that in the NO scenario, while bottom salinity differences between the DPD and the NO scenarios can be as high as \( \sim 3 \) in the Barataria Bay. On the other hand, the maximum salinity differences between the MBD and NO scenarios for both surface and bottom exhibited a great decrease, \( \sim 12 \), in the Barataria Bay, with a larger domain at the bottom compared with the surface. The DPD could impact residence times of Little Lake and Barataria Bay. In the MBD scenario, to the contrary, residence times of a large portion of the estuary could be greatly affected due to changes in residual currents and water exchanges.

In chapter 3, the numerical model was configured to simulate temporal and spatial variability of the lateral circulation in the Barataria Pass, over a 25.6-hour diurnal tidal cycle. Model performance was assessed against observational data. The density distribution within any estuary is a result of both advective and mixing processes. In Barataria Pass, the turbulent mixing is closely related to the magnitude of ebb/flood current and the strength of the tidal bottom
boundary layer. Characteristics of horizontal advection processes in the inlet are that maximum flood currents are located at the central part of the deep channel for a large part of the flood period. This differential advection (Nunes and Simpson, 1985), when acting upon the along-channel density gradient, produces a distinct density difference between the shoal and channel waters. In addition, the advection of Mississippi River water to the eastern channel during part of the flood period further enhances the density difference. On the contrary, maximum ebb currents swing between the western slope and central surface of the channel during the ebb. When maximum ebb flows are at the western slope, the differential advection mechanism does not work. When they go back to the channel center, the salinity contour lines are mostly horizontal due to weak vertical turbulence mixing. Thus, both situations are not favorable to produce an extreme density near the middle of the channel. During flood period, when density distribution is high near the channel center and low at both shoals, the horizontal pressure gradient drives a lateral circulation with two counter-rotating cells and surface or near surface convergence. This result from the Barataria Pass is similar to that reported by Nunes and Simpson (1985). However, detailed analysis of momentum equations indicates that, in addition to the pressure gradient and vertical stress divergence, nonlinear advection and horizontal stress divergence are also important terms. During ebb period, the lateral circulation is mostly eastward for the whole water column and persisting for almost the whole period. The surface divergence suggested by the differential advection mechanism is either non-existent or lasting for very short period. The main momentum balance across most of the transect is between the along-channel advection of cross-channel momentum and pressure gradient. In addition, the sectional averaged lateral velocity magnitude during ebb is comparable to that during flood, which is different from the idealized numerical experiment (Lerczak and Geyer, 2004).
To more closely examine estuarine-shelf exchange, the circulation and particle transport through the Barataria Pass, and over the continental shelf were simulated using a three-dimensional hydrodynamic and Lagrangian particle tracking model in chapter 4. Predicted large-scale circulation on the continental shelf of the Northern Gulf of Mexico was consistent with previous reports in that it was primarily wind-driven. Outflows through the inlet advected fresher estuarine water onto the saline shelf water to form a radially spreading estuarine plume in the coastal ocean. The evolution of the estuarine plume within a diurnal tidal cycle was described in detail. Wind-driven coastal circulation determined the subtidal variations of the plume when the upstream freshwater discharge rate was almost constant. Cross-shore winds determined the landward retardation or seaward spreading of the plume, while along-shore winds lead to asymmetry in the along-shore plume geometry. Internal hydraulic jumps were also identified on the plume interface. Under landward winds, surface passive particles released at the beginning of flood tides could be transported ~ 12 km landward of the inlet, and could stay inside the estuary during the next ebb period. On the contrast, under seaward winds the same group of surface passive particles was expelled offshore during the next ebb tide. Particles released at different layers showed that those advected by mid-depth currents entered the inlet much more easily than those released at surface and near bottom.

This dissertation covers broad topics in the study of estuary-inlet-shelf system but has not given too many mechanistic interpretations. However, there are many specific aspects of this research that deserve detailed future study, including, mixing and dispersion of the plume frontal zone, estuarine plume-related internal wave generation, and mechanisms for coastal upwelling/downwelling.
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APPENDIX. DECOMPOSITION OF VECTORS INTO ALONG- AND CROSS-CHANNEL DIRECTIONS

In $x$-$y$-$\sigma$ coordinates, where $x$-direction is defined to the east and $y$-direction to the north.

The FVCOM $x$- and $y$-axis 3-D momentum equations are written as:

$$
\frac{\partial u}{\partial t} + \frac{\partial u D}{\partial x} - \frac{\partial w D}{\partial y} - \frac{\partial \omega D}{\partial \sigma} + \frac{f v D}{Cy} - \frac{g D}{\rho_0} \left[ \frac{\partial}{\partial x} \left( \int_0^\sigma d \sigma' + \sigma \rho \frac{\partial D}{\partial x} \right) \right] + \\
\frac{1}{\rho_0} \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial u}{\partial \sigma} \right) + F_x_{HVISY} = 0
$$

(A1)

$$
\frac{\partial v}{\partial t} + \frac{\partial u D}{\partial y} - \frac{\partial w D}{\partial x} - \frac{\partial \omega D}{\partial \sigma} - \frac{f u D}{Cy} - \frac{g D}{\rho_0} \left[ \frac{\partial}{\partial y} \left( \int_0^\sigma d \sigma' + \sigma \rho \frac{\partial D}{\partial y} \right) \right] + \\
\frac{1}{\rho_0} \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial v}{\partial \sigma} \right) + F_y_{HVISY} = 0
$$

(A2)

where $(u, v)$ are the velocity components in the $(x, y)$ directions.

In order to quantify along- and cross-channel momentum balance, we chose the along-channel direction ($y'$) to be aligned with the channel. This value is positive when pointing into the estuary. The cross-channel direction ($x'$) is defined as positive when pointing to the eastern boundary (Fig. A1). The relationship between $(u', v')$ and $(u, v)$, $(x', y')$ and $(x, y)$ are as follows:

$$
\begin{align*}
    u' &= u \cos \theta + v \sin \theta \\
    v' &= -u \sin \theta + v \cos \theta \\
    x' &= (x - x_0) \cos \theta + (y - y_0) \sin \theta
\end{align*}
$$

(A3)  
(A4)  
(A5)
\[ y' = -(x - x_0) + (y - y_0) \cos \theta \] (A6)

where \( \theta \) is the angle between the \( x' \)-direction (cross-channel) and the \( x \)-direction.

Figure A.1. Illustration of \( x - y \) coordinate transformed to \( x' - y' \) coordinate.

In \( x' - y' \) coordinates, the momentum equations are written as:

\[
\frac{\partial u_D}{\partial t} = -\frac{\partial u_D^2}{\partial x_I} - \frac{\partial u_D v_D}{\partial y_I} - \frac{\partial u_D \omega}{\partial \sigma} + f u_D' - g D \frac{\partial z}{\partial x_I} \frac{D}{\rho_0} \left[ \frac{\partial}{\partial x_I} \left( D \int_0^0 d \sigma' + \sigma \rho \frac{\partial D}{\partial x_I} \right) \right] + \\
\frac{1}{D} \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial u_I}{\partial \sigma} \right) + F_{x_I}^{\text{VISC}}  
\] (A7)

\[
\frac{\partial v_D}{\partial t} = -\frac{\partial u_D v_D}{\partial x_I} - \frac{\partial v_D^2}{\partial y_I} - \frac{\partial v_D \omega}{\partial \sigma} - f v_D' - g D \frac{\partial z}{\partial y_I} \frac{D}{\rho_0} \left[ \frac{\partial}{\partial y_I} \left( D \int_0^0 d \sigma' + \sigma \rho \frac{\partial D}{\partial y_I} \right) \right] + \\
\frac{1}{D} \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial v_I}{\partial \sigma} \right) + F_{y_I}^{\text{VISC}}  
\] (A8)

To project the momentum equations into the cross- and along-channel directions, we treated each term in the momentum equations as a vector in the \((x, y)\) direction and then applied
the same decomposition as eqs. (A3, 4). Thus, terms in $x'-y'$ coordinates can be calculated by corresponding terms in $x-y$ coordinates as follows:

\[
DUDT' = \frac{\partial u'D}{\partial t} = \frac{\partial (ucos\theta + vsin\theta)D}{\partial t} = \frac{\partial uD}{\partial t}cos\theta + \frac{\partial vD}{\partial t}sin\theta
\]

\[
= DUDT cos\theta + DVDT sin\theta
\]

\[
DVDT' = \frac{\partial v'D}{\partial t} = \frac{\partial (-usin\theta + ucos\theta)D}{\partial t} = -\frac{\partial uD}{\partial t}sin\theta + \frac{\partial vD}{\partial t}cos\theta
\]

\[
= -DUDT sin\theta + DVDT cos\theta
\]

Terms ADVUX', ADVVX', ADVUY', ADVVY', ADVWX', ADVWY', CORX', CORY', DPBPX', DPBPY', DPBCX', DPBCY', VVISCX', VVISCY', HVISCX', and HVISCY' can be calculated with the same method. While ADVUX', ADVVX', ADVUY' and ADVVY' should be calculated by the finite volume difference. For example, ADVUX' can be calculated as:

\[
\int \int \frac{\partial u'^2 D}{\partial x'} dx'dy' = \oint u'u'Ddy' = \oint UIJ' \times \left(\frac{uij1'}{uij2'}\right) \times DIJ \times dy'
\]

(A9)

where $UIJ$, $UIJ1'$ and $UIJ2'$ are shown in Fig. A2.

Figure A.2. Illustration of local coordinate used to calculate the horizontal advection terms.
With eqs. (A3-A6), we have:

\[ dy' = -\sin \theta + \cos \theta dy \]  
(A10)

\[ UIJ' = UIJ \cos \theta + VIJ \sin \theta \]  
(A11)

\[ UIJ'_2 = UIJ'_1 \times \cos \theta + VIJ'_1 \times \sin \theta \]
(A12)

Substituting eqs. (A10-A12) into eq. (A9),

\[ ADVUX' = \frac{\partial u'^2 D}{\partial y'} \]

\[ = \frac{\partial u^2 D}{\partial y} \sin \theta \cos^2 \theta + \frac{\partial w D}{\partial y} \sin^2 \theta \cos \theta + \frac{\partial v D}{\partial y} \sin^2 \theta \cos \theta + \frac{\partial v^2 D}{\partial y} \sin^3 \theta \]

\[ + \frac{\partial u^2 D}{\partial x} \cos^3 \theta + \frac{\partial u v D}{\partial x} \sin \theta \cos^2 \theta + \frac{\partial v u D}{\partial x} \sin \theta \cos^2 \theta + \frac{\partial v^2 D}{\partial x} \sin^2 \theta \cos \theta \]

With the same method, \(ADVUX', ADVUY'\) and \(ADVYY'\) are given as:

\[ ADVUX' = \frac{\partial u'v'D}{\partial y'} \]

\[ = -\frac{\partial u^2 D}{\partial y} \sin \theta \cos^2 \theta - \frac{\partial u v D}{\partial y} \sin^2 \theta \cos \theta + \frac{\partial v u D}{\partial y} \cos^3 \theta + \frac{\partial v^2 D}{\partial y} \sin \theta \cos^2 \theta \]

\[ + \frac{\partial u^2 D}{\partial x} \sin^2 \theta \cos \theta + \frac{\partial u v D}{\partial x} \sin^3 \theta - \frac{\partial v u D}{\partial x} \sin \theta \cos^2 \theta - \frac{\partial v^2 D}{\partial x} \sin^2 \theta \cos \theta \]

\[ ADVUY' = \frac{\partial u'v'D}{\partial x'} \]

\[ = -\frac{\partial u^2 D}{\partial y} \sin^2 \theta \cos \theta + \frac{\partial u v D}{\partial y} \sin \theta \cos^2 \theta - \frac{\partial v u D}{\partial y} \sin^3 \theta + \frac{\partial v^2 D}{\partial y} \sin^2 \theta \cos \theta \]

\[ - \frac{\partial u^2 D}{\partial x} \sin \theta \cos^2 \theta + \frac{\partial u v D}{\partial x} \cos^3 \theta - \frac{\partial v u D}{\partial x} \sin^2 \theta \cos \theta + \frac{\partial v^2 D}{\partial x} \sin \theta \cos^2 \theta \]
\[ \text{ADVY}' = \frac{\partial v'^2 D}{\partial y'} \]

\[ = \frac{\partial u^2 D}{\partial y} \sin^2 \theta \cos \theta - \frac{\partial uvD}{\partial y} \sin \theta \cos^2 \theta - \frac{\partial vuD}{\partial y} \sin \theta \cos^2 \theta + \frac{\partial v^2 D}{\partial y} \cos^3 \theta \]

\[ - \frac{\partial u^2 D}{\partial x} \sin^3 \theta + \frac{\partial uvD}{\partial x} \sin^2 \theta \cos \theta + \frac{\partial vuD}{\partial x} \sin^2 \theta \cos \theta - \frac{\partial v^2 D}{\partial x} \sin \theta \cos^2 \theta \]
VITA

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