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NUMERICAL AND FIELD STUDY OF TIDAL AND SUBTIDAL DYNAMICS IN A BAR-BUILT ESTUARY: BARATARIA BAY, GULF OF MEXICO

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 & Coastal Sciences

by

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B.S., University of Mohaghegh Ardabili, 2011
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August 2021
To my wife, Afsaneh who has a heart the size of an ocean
and I found myself in its hidden depths
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ABSTRACT

This study investigated tidal and subtidal dynamics of water level, currents, and suspended sediment concentration (SSC) in Barataria Bay, a shallow bar-built estuary of the Northern Gulf of Mexico. First, the local and remote wind forcing contribution on subtidal water level and current variability were examined using three different methods: (i) statistical analysis of observed data, (ii) an analytical model and (iii) a 2-D barotropic numerical model. Results suggested that the remote and local wind effects were equally important at the bay mouth, however local winds were the dominant forcing driver inside the bay. The amplitudes of subtidal fluctuations induced by local winds were twice as large as the one caused by remote winds. This finding differs from those found in the existing literature, notably for Breton Sound and Lake Pontchartrain, where remote wind effect has been reported to be dominant. These differences are attributed to the different geomorphological features of the estuaries. Furthermore, the seasonality of the SSC in Barataria Pass was explored as the offshore environment transitioned from a period of high cold front activity and low river discharge to a period of low cold front activity and high river discharge. The SSC dynamics during the winter was mainly forced by resuspension in response to the cold front winds. During the spring, the average SSC (0.23 g l\(^{-1}\)) was significantly higher than winter (0.15 g l\(^{-1}\)) because of the strong offshore influence of the Mississippi River plume. Finally, tidal response to the relative sea level rise and marsh accretion was investigated. Contrary to previous modeling analyses in other estuaries suggesting that flooding of the low-lying land with sea level rise would increase frictional effects and thus reduce tidal range, this study suggested that tidal range in a choked tidal system like Barataria Bay increases even when accompanied by extensive land inundation. This occurs because in Barataria Bay the channel conveyance effects are larger than the frictional effects of the low-lying areas. In the lower and the middle bay, the largest increase in tidal range occurred when the marsh area was assumed to keep pace with relative sea Level rise. However, in the upper bay the largest increase in tidal range occurred when no accretion was assumed. In addition, relative sea level rise caused tidal amplification at the head of the estuary. A detailed momentum balance analysis indicated that sea level rise shifts tidal wave from a dissipative regime to a more progressive wave, which is more likely to be amplified.
CHAPTER 1. INTRODUCTION

1.1 Background and Research Objectives

Estuaries are very productive and fertile ecosystems in the world. They serve as critical natural habitats, providing shelter and breeding grounds to a wide variety of species that are valuable biologically, commercially, culturally, and recreationally. They are also economically indispensable as they are links between land and sea, providing access to inland waterways and harbors. The fact that many large cities in the world are located on estuaries is a clear evidence that estuaries are the preferred site for human settlement and human being is totally dependent on estuaries (Ross, 1995). Besides serving as important habitat for wildlife and their economic benefits, estuaries perform environmental services. They act as protective buffers between the land and the ocean dissipating storm surges and absorbing flood waters. Salt marshes and mangrove forests associated with estuaries act like natural filters soaking up sediments and pollutants such as excess nutrients, heavy metals, pesticides, and herbicides out of the water.

Nowadays, estuaries around the world are facing serious challenges owing to various anthropogenic activities and natural changes. Bricker et al. (1999), for example, examined 138 estuaries in the US and reported that 60% of them displayed modest to severe nutrient enrichment problems. In addition, estuaries are facing serious impacts from reduced sediment delivery, pollution impacts due to a wide variety of chemical contaminants being released, introduced species, and with climate change, they are confronting serious impacts from sea level rise.

Detailed examination of estuarine hydrodynamics improves our understanding of their ecosystem dynamics, material transport and helps us predict their possible changes in the future. Water exchange in estuaries is subject to different forcing mechanisms including winds, astronomical tides, river discharge, and buoyancy forces, and is modulated by earth’s rotation and bathymetry. These forces induce fluctuations in water levels ranging from a few hours to days and months. Two of the most important aspects of estuarine dynamics are tidal and the tidally averaged or tidal residual circulation. Tidal processes play a significant role in the exchange of water, sediments, pollutants, nutrients, and salt, between estuaries and the coastal ocean. Additionally, residual dynamics, also known as subtidal dynamics, over days to months can also play a major role in regulating estuarine circulation. Subtidal dynamics are especially important in microtidal estuaries where astronomical tidal ranges are relatively small (Snedden et al., 2007).

Coastal Louisiana influenced by the largest river in North America, the Mississippi River, comprises many microtidal estuaries that are undergoing profound deterioration and shrinking as a result of human activities and natural changes (Fisk, 1952; Roberts, 1997; Coleman et al., 1998). Located between the Mississippi River and Bayou Lafourche, Barataria Estuary is at the heart of these rapid changes in Louisiana (Figure 1-1). Barataria Estuary, also known as Barataria Bay, is a bar-built Mississippi River estuary and is about 100 km long from the mouth to its head in Lac Des Allemands. The lower estuary encompasses large open water bodies, and the upper estuary mostly comprises divided lakes and water bodies connected with bayous and channels to the lower estuary. Barataria Basin is very shallow and has an average depth of about 2 m.
Circulation in Barataria Bay is more complex than in a typical estuary. The major freshwater input for Barataria Bay is a man-made freshwater diversion, the Davis Pond which with the maximum designed capacity of ~300 m$^3$ s$^{-1}$. Other sources of fresh water in Barataria are from rainfall and stream runoff. In addition, the Gulf Intracoastal Waterway (GIWW) delivers on average 50 m$^3$ s$^{-1}$ of water coming from the Atchafalaya River into Barataria basin (Swarzenski and Perrien 2015) and acts as an unintended river diversion. GIWW is a ~100 m wide channel with an average depth of ~5 m, which was constructed for commercial navigation from Florida to Texas. It is estimated that GIWW annually delivers on average 0.21 Tg of
sediment to mid-Barataria. This sediment is transported for more than 100 km before reaching Barataria and comes mostly from the Atchafalaya River (Mariotti et al., 2021).

Subtidal dynamics in the northern Gulf of Mexico are influenced by cold fronts. A cold front occurs when a mass of relatively colder air moves into where warmer air is present. The colder and drier air lifts the warmer and moisture air making cloud formation. This often causes a line of showers and thunderstorms when enough moisture is present. These cold fronts are associated with a clockwise wind field that shifts from southerly during the prefrontal stage to westerly and then northerly during the frontal stage. Cold fronts are more common between October and April, when strong and frequent northerly winds affect the Northern Gulf of Mexico every 3 to 8 days (Chuang & Wiseman, 1983; Stone & Wang, 1999). Cold fronts are categorized into two different types: eastward migrating cyclones and southward migrating arctic surges (Roberts et al., 1987).

Direct action of cold front winds on estuarine surface affects estuarine water level and currents and therefore causes significant changes in flushing times of estuary. Feng and Li (2010), for example, studied three Louisiana estuaries and found out that a strong cold front can flush out more than 40% of the estuarine water in less than 40 hr. They found that northerly winds are the most important force driving estuarine-shelf exchanges. Li et al., (2011) partitioned the observed water level setup in Atchafalaya Bay and found that 50% of the setup was due to winds, 25% was caused by atmospheric pressure and 25% was induced by wave setup.

Wind-driven subtidal variability in coastal embayments and estuaries cab be divided into two major parts: remote wind effects and local wind effects. Local wind effect is caused by direct action of the wind on the estuarine water surface and thus is mainly linked to along-estuary winds. Remote wind effect is caused by cross-estuary winds inducing perturbations in the coastal ocean in response to the coastal Ekman transport. The resulting variations can travel inside and induce water level fluctuations within the estuary (Garvine, 1985; Janzen & Wong, 2002; Wong & Wilson, 1984).

Subtidal dynamics of relatively deep and partially mixed coastal plain estuaries is well studied (e.g., Garvine, 1985; Bosley & Hess, 2001; Henrie & Valle-Levinson, 2014; Wong & Garvine, 1984), as well as fjord estuaries (e.g., Mofjeld, 1992) and tectonic estuaries (e.g., Walter & Gartner, 1985; Walters, 1982). These studies mostly indicated the domination of remote winds over local winds in governing subtidal variations. Wang and Elliott (1978), for example found that subtidal variations in Chesapeake Bay are mostly the result of coastal sea level variations propagated into the estuary. Likewise, subtidal water level changes in San Francisco Bay, a tectonic estuary, are primarily governed by remote wind effects (Walters, 1982; Walter & Gartner, 1985). In Delaware Estuary, fluctuations caused by coastal Ekman transport are dominant for all subtidal frequencies (Wong and Garvine 1984). Subtidal water level variations in Puget Sound are strongly coupled to the coastal sea level fluctuations especially in the southern reaches of this fjord estuary (Mofjeld, 1992).

However, subtidal dynamics of shallow bar-built estuaries have received less attention compared to other estuarine types. Bar-built estuaries, such as Barataria Bay, are restricted-mouth coastal embayments and they are commonly found along Gulf Coasts of the United States. Bar-built estuaries are usually microtidal and therefore subtidal variability could be of greater or comparable magnitude than astronomic tides. In addition, water exchanges in bar-built estuaries are limited due to the presence of the barrier islands separating the estuary from the coastal ocean. Therefore, the dominance of remote wind effects over local wind effects may not be evident for these kinds of estuaries. Therefore, my first research objective was to quantify the
relative contribution of remote and local wind effects on subtidal water level and current dynamics in Barataria Bay during cold front season.

Offshore freshwater and sediment sources (i.e., neighboring river) can significantly impact sediment dynamics within an estuary (e.g., Yamada & Kosro, 2010; McConnaughey et al., 1994). Southwest Pass which is the main outlet of the Mississippi River is located 60 km to the south of Barataria Bay and exports 45–67% of the Mississippi River discharge (Allison et al., 2012). Southwest path is a ~700 m wide channel and experiences supercritical flow around the mouth of the outlet with velocities more than 2 m s\(^{-1}\) (Sorourian et al., 2020). A clockwise gyre in the Louisiana Bight could occasionally carry the plume of the Mississippi River northward towards Barataria Bay (Rouse and Coleman, 1976; Wiseman et al., 1976; Walker et al., 2005). In addition, West Bay Diversion and various nearby crevasses along the lower Mississippi River discharge Mississippi water onto the shelf. A portion of these sediment laden waters is transported northwestward with the help of the Louisiana Coastal Current (Walker, 1996). Consequently, Mississippi river plume may enter the estuary through its tidal inlets. The average salinity on the eastern side of the Barataria Pass is lower as a result of the Mississippi river plume intrusion from the east side (Li et al., 2009). Contrary to the engineered structures like river diversions and GIWW, the offshore influence of the Mississippi River plume on salinity and sediment dynamics in Barataria Bay is relatively unknown and needs to be investigated. My second research objective was to explore how suspended sediment concentration in Barataria Bay vary seasonally, as the region transitions from a period of high cold front activity and low river discharge to a period of low cold front activity and high river discharge.

Beyond subtidal processes, astronomical tides are arguably the most important factor determining the hydrodynamics of estuaries and coastal embayments. Tidal dynamics in estuaries are expected to be affected by sea level rise. Some of the highest relative sea level rise (RSLR = eustatic sea level rise (ESLR) + subsidence) rates in the continental USA occurs in Coastal Louisiana (0.92 cm yr\(^{-1}\) Grand Isle tide gage: 1947-2006; Miner, 2007) as compared to ~1.5 mm yr\(^{-1}\) of ESLR. As a result, Barataria Bay has experienced substantial wetland loss in recent decades (16.9 km\(^2\) yr\(^{-1}\), from 1935-2000; FitzGerald et al., 2007).

Tidal range response to RSLR is complex and has been showed to either increase or decrease. Increasing tidal ranges with sea level rise is reported in many estuaries (Holleman and Stacey 2014; Dominicis et al., 2017; Lee et al., 2017). Lee et al. (2017) investigated how sea level rise may impact tides in Chesapeake Bay and Delaware Bay. They found similar responses in both estuaries, such that when the low-lying land is prevented from flooding, tidal range increases and when it is allowed to become permanently inundated by higher sea level, tidal range decreases. Similarly, Holleman & Stacey (2014) studied tidal changes in San Francisco Bay under future sea level rise scenarios using a numerical model. They found that flooding of low-lying regions introduces frictional areas that serve as energy sinks for tides, thus decreasing the tidal range under higher sea levels. Despite of these similarities, tidal response to sea level rise varies among different estuaries depending on the estuarine characteristics (Passeri et al., 2016; Du et al., 2018).

Barataria Bay has unique features that differentiate it from other estuaries. It features a complex geometry involving many bayous, lakes, channels and extensive tidally influenced marshes. Additionally, Barataria Bay is a tidally choked estuary. Tidal choking occurs when a relatively large surface area is connected to a narrow, long, or shallow channel causing significant reductions in tidal range. Therefore, tides in Barataria Bay are dissipated heavily so that they are attenuated by 68% at Lafitte located at the middle of the estuary (Figure 1-1; Byrne
et al., 1976; Conner et al., 1987). The combined effects of RSLR and potentially increasing tidal ranges will have pervasive effects on Barataria Bay. Thus, as sea level rises, it is important to quantify how tidal dynamics will be affected. My third research objective was to investigate how the combined effects of relative sea level rise and marsh accretion will change tidal dynamics in Barataria Bay.

1.2 Synopsis of Chapters

My overarching research objectives were addressed in three chapters. In Chapter 2, I investigated the effects of local and remote wind forcings on the subtidal water level and current variability in Barataria Bay. Three different methods were employed: (i) statistical analysis based on field measurements, (ii) an analytical model, and (iii) a 2-D barotropic numerical model based on the Finite Volume Community Ocean Model (FVCOM). In Chapter 3, I studied the seasonal variability in the suspended sediment concentration within the main inlet of Barataria Bay, as the offshore environment transitioned from a period of high cold front activity and low river discharge to a period of low cold front activity and high river discharge. This study was based on two rounds of field observations carried out during the winter of 2017-2018 and spring of 2018. In chapter 4, I explored tidal change in response to the relative sea level rise and marsh accretion in Barataria Bay using a finite volume community ocean model (FVCOM). In chapter 5, I presented a summary of the results and conclusions from the entire study.
CHAPTER 2. SUBTIDAL WATER LEVEL AND CURRENT VARIABILITY IN A BAR-BUILT ESTUARY DURING COLD FRONT SEASON: BARATARIA BAY, GULF OF MEXICO

2.1 Introduction

Processes at subtidal time scales (i.e., days to months) can play an important role in governing estuarine dynamics (Pritchard, 1955; Weisberg, 1976). Wind-driven subtidal variability in estuaries and coastal embayments is mainly associated with two components: local wind effects and remote wind effects. The former is caused by wind stresses directly acting on the estuarine water surface and is mainly associated with along-estuary winds. The latter is caused by cross-estuary winds imposing the water level in the coastal ocean in response to coastal Ekman transport (Garvine, 1985; Janzen & Wong, 2002; Wong & Wilson, 1984).

Studies of subtidal circulation have often focused on narrow, relatively deep and partially mixed coastal plain estuaries (Bosley & Hess, 2001; Garvine, 1985; Henrie & Valle-Levinson, 2014; Wang & Elliott, 1978; Wong & Garvine, 1984), tectonic estuaries (Walters, 1982; Walter & Gartner, 1985), and fjord estuaries (Mofjeld, 1992). Most of these studies pointed to the dominance of remote wind effects over local wind effects in controlling subtidal variations. Several examples support this idea; for example, Wang and Elliott (1978) found that the dominant water level variations in Chesapeake Bay were the result of up-bay propagation of coastal sea level variations, which in turn were generated by coastal Ekman transport associated with alongshore winds. Likewise, Wong and Garvine (1984) found that remote wind effects caused by coastal Ekman transport were dominant within the Delaware Estuary for all subtidal frequencies, while local wind forcing was insignificant. Henrie and Valle-Levinson (2014) showed that subtidal water levels generated from the coastal ocean in St. Johns River estuary can propagate up to 145 km upstream of the inlet where the tidal waves exhibit attenuation, slight amplification, and then again strong attenuation when moving from the inlet to the upstream end. Low-frequency sea level changes in San Francisco Bay, a tectonic estuary, have shown to be primarily controlled by remote coastal sea level variations (Walters, 1982; Walter & Gartner, 1985). Further, observations in Puget Sound showed that subtidal water level variations in the southern reaches of this fjord estuary were strongly coupled to the coastal sea level fluctuations (Mofjeld, 1992). All of these results are consistent with a simple barotropic model (Garvine, 1985) showing that, due to the relative shortness of most estuaries compared to subtidal wavelengths, the remote wind effect is dominant for both sea level and barotropic current fluctuations.

Subtidal dynamics in shallow bar-built estuaries, the third major estuarine type commonly found along the southeast and Gulf Coasts of the United States, have received less attention than in the other types of estuaries. In bar-built estuaries the subtidal variability could be of comparable or greater magnitude than astronomic tides. This happens because even small tidal amplitudes are largely attenuated due to their highly frictional landscapes and complex geometries (Conner et al., 1987; Copeland et al., 1968). For example, Copeland et al. (1968) indicated that the wind effects outweighed the tidal effects in controlling the surface dynamics of

Laguna Madre, a bar-built estuary along the Texas coasts. These dynamics are highly relevant for the northern Gulf of Mexico, where astronomic tidal ranges are small (~0.3 m).

An open question regarding subtidal variability in the northern Gulf of Mexico is whether that is mainly associated with the remote or local forcing and whether the effects are modulated by the geometry of the estuary. Snedden et al. (2007) studied the relative role of local and remote forcings on water level variability in Breton Sound, an estuary with a wide connection to the Gulf of Mexico and found that the effect of local wind stress over the estuarine surface was minimal due to the lack of significant fetch. Huang and Li (2017) showed that in Lake Pontchartrain, an ~4-m deep estuary with a small marsh cover, remote wind is the main contributor to the total water level change, but the surface slopes across the estuary are mostly governed by local winds. Both above-mentioned studies point to the significant role played by remote winds as opposed to local winds, but the limited number of studies prevents generalization of the results to estuaries with different geometries.

In order to study subtidal dynamics in the northern Gulf of Mexico, it is necessary to consider its peculiar meteorology, which is dominated by cold fronts. Cold fronts are more pronounced between October and April, when frequent and strong northerly wind events affect the region every 3 to 8 days (Chuang & Wiseman, 1983; Stone & Wang, 1999). Roberts et al. (1987) described two different types of cold fronts: eastward migrating cyclones and southward migrating arctic surges. Water level setup and setdown associated with cold fronts causes significant changes in estuarine residence times (Feng & Li, 2010; Moeller et al., 1993). For example, Feng and Li (2010) studied subtidal water exchanges in three Louisiana estuaries and found that strong cold fronts can flush out more than 40% of the estuarine volumes in less than 40 hr. They suggested that both cross-estuary and along-estuary winds affect estuarine-shelf water exchanges, with winds coming from the northeast being the most influential. Partitioning of the observed water level setup in Atchafalaya Bay revealed that 50% of the setup was wind-induced, 25% was induced by atmospheric pressure, while 25% was due to wave setup (Li et al., 2011).

This study examines the effects of local and remote wind forcings during cold front season on the subtidal water level and current variability in Barataria Bay, a bar-built Mississippi River Estuary. Barataria Bay is characterized by the limited access to the shelf, large marsh cover, and a depth of only 2 m. Barataria Bay is also the site of major coastal restoration projects, where two large-scale river diversions are currently under considerations (Coastal Protection and Restoration Authority, 2017). Those river diversions will likely strongly affect water levels in the vicinity of diversion sites, and so understanding the wind-driven water level dynamics in this estuary is of fundamental importance to correctly assess the potential effects of proposed river diversions. Three different methods were employed in the analysis: (1) statistical analysis based on measurements of water levels and currents, (2) an analytical model with idealized wind field, and (3) a barotropic numerical circulation model based on the Finite Volume Coastal Ocean Model (FVCOM).

2.2 Study Site

Barataria Bay is a bar-built Mississippi River estuary located in south Louisiana, along the northcentral Gulf Coast of the United States (Figure 2-1). The estuary is about 52 km long from Bayou Perot to its mouth at Barataria Pass. The lower 30 km of the estuary encompasses over 80% of the overall surface area of the estuary and has an average depth of 2 m. Barataria
Bay has four inlets, with Barataria Pass being the main inlet, accounting for ~66% of total water exchange (Marmer, 1948). A relatively deep shipping channel (~4 m) runs from the Barataria Pass along the axis of the estuary and extends to the Gulf Intracoastal Waterway. The width of the estuary is about 50 km at its widest point near the mouth. The estuary is oriented ~340°T (clockwise from true north), and estuarine axis relative to the coastline is ~90°T. The adjacent continental shelf is 80 km wide with a gentle bottom slope on the order of 0.1%. Southwest Pass, located 60 km to the south of Barataria Bay, is the main outlet of the Mississippi River (Allison et al., 2012; Li, White, et al., 2011). A clockwise gyre in the Louisiana Bight carries much of the Mississippi River fresh water, sediments, and nutrients back to the shoreline. Consequently, circulation in this confluence region of Barataria Bay, Mississippi River, and adjacent shelf, is much more complicated than in a typical estuary.

Figure 2-1 Study area and locations of monitoring stations. BB1 and BB3 denote the stations in Barataria Pass and Bay Jimmy, respectively, where the water levels were measured during this study. BB2, BB4, BB5, and BB6 denote stations where the water levels were collected by the United States Geological Survey National Water Information System. Currents were measured in station BB1. BB7 denotes the location of United States Geological Survey station where the wind measurements were carried out. The positive direction of along-estuary (al) and cross-estuary (cr) winds are represented by blue arrows.

Barataria Bay is a micro tidal estuary, with $K_1$ and $O_1$ being the dominant tidal constituents, whose amplitudes at Grand Isle are about 11 cm. The maximum astronomic tidal range at Grand Isle is 62 cm, and tides are attenuated by 68% at Lafitte (Byrne et al., 1976; Conner et al., 1987). Subtidal variations can be as high as 100 cm during cold front passage events (Walker & Hammack, 2000) and are, therefore, important controlling mechanism governing water level and movement of fresh water, sediments, and nutrients in Barataria Bay.
2.3 Methods

2.3.1 Data Acquisition

Primary data used in this study include water level, currents, and winds. The water level data were measured by pressure transducers at two stations, one in Barataria Pass (BB1) and another one in Bay Jimmy (BB3) from 7 December 2016 until 5 March 2017. In addition, water level data were simultaneously collected at four representative stations (BB2, BB4, BB5, and BB6) included in the United States Geological Survey (USGS) National Water Information system (HTTPS://WATERDATA.USGS.GOV/NWIS). The locations of stations BB2, BB4, BB5, and BB6 correspond to USGS site numbers 291929089562600, 29285909004000, 292800090060000, and 07380335, respectively (Figure 2-1). Currents were measured using an Acoustic Doppler Velocimeter in Barataria Pass (station BB1) where the water depth is 3 m (Figure 2-1). The Acoustic Doppler Velocimeter was placed at 0.4 m above the bottom. Water levels and currents were burst sampled at 4 Hz for 5 min every hour.

Barataria Pass is the main inlet of the estuary where the bulk volume transport occurs, which allowed us to examine the effects of local and remote forcing on surface currents. Cold fronts are large-scale weather systems for which the spatial variation in wind speed is negligible. The USGS provided 30-min wind and atmospheric pressure observations from a meteorological station north of Barataria Bay (BB7: USGS [07380251]). Wind data from this station were shown to adequately represent wind stress within the interior of the estuary (Mariotti et al., 2018). Prior to determining the influence of winds on water level variability, the water levels were adjusted based on pressure measurements to remove the effects of atmospheric pressure fluctuations (i.e., the inverted barometer effect).

2.3.2 Cold Front Characterization

During the winter months, the dominant synoptic features in this region are the cold fronts, which typically pass through southern Louisiana with a recurrence interval of 3–8 days. Associated with these frontal passages is a clockwise wind field which shift from southerly during the prefrontal phase to westerly and then northerly during the frontal phase. The 3-hr surface analyses maps produced by National Oceanic and Atmospheric Administration (NOAA)'s Hydrometeorological Prediction Center (HTTPS://WWW.WPC.NCEP.NOAA.GOV/ARCHIVES/WEB_PAGES/SFC/SFC_ARCHIVE.php) were used to identify cold fronts, and meteorological measurements from USGS northern Barataria Bay station 07380251 were used to pinpoint the timing of the frontal events. A total of 21 cold fronts were identified during the observation period. Of these, four cold fronts did not possess a pressure trough, a postfrontal temperature decrease or a shift in the wind from southerly to northerly and were therefore eliminated from the analysis. Of the remaining 17 cold fronts, 13 were migrating cyclones, and four were arctic surges (Table 2-1). Migrating cyclones move eastward while arctic surges move southward (Roberts et al., 1987).
Table 2-1 Winter Frontal Passages From 7 December 2016 to 5 March 2017

<table>
<thead>
<tr>
<th>Start Time</th>
<th>End Time</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 December 7, 2016, 09:00</td>
<td>December 11, 2016, 17:30</td>
<td>MC</td>
</tr>
<tr>
<td>2 December 14, 2016, 09:00</td>
<td>December 16, 2016, 16:30</td>
<td>AS</td>
</tr>
<tr>
<td>3 December 18, 2016, 15:00</td>
<td>December 22, 2016, 06:00</td>
<td>MC</td>
</tr>
<tr>
<td>4 December 22, 2016, 12:00</td>
<td>December 23, 2016, 15:00</td>
<td>MC</td>
</tr>
<tr>
<td>5 December 29, 2016, 15:00</td>
<td>December 31, 2016, 05:30</td>
<td>MC</td>
</tr>
<tr>
<td>6 January 3, 2017, 06:00</td>
<td>January 3, 2017, 15:00</td>
<td>MC</td>
</tr>
<tr>
<td>7 January 4, 2017, 09:00</td>
<td>January 5, 2017, 07:06</td>
<td>MC</td>
</tr>
<tr>
<td>8 January 6, 2017, 03:00</td>
<td>January 9, 2017, 13:30</td>
<td>MC</td>
</tr>
<tr>
<td>9 January 22, 2017, 12:00</td>
<td>January 24, 2017, 15:00</td>
<td>MC</td>
</tr>
<tr>
<td>10 January 26, 2017, 06:00</td>
<td>January 29, 2017, 17:30</td>
<td>MC</td>
</tr>
<tr>
<td>11 February 3, 2017, 09:00</td>
<td>February 5, 2017, 09:00</td>
<td>AS</td>
</tr>
<tr>
<td>12 February 9, 2017, 09:00</td>
<td>February 10, 2017, 14:00</td>
<td>AS</td>
</tr>
<tr>
<td>13 February 13, 2017, 09:00</td>
<td>February 14, 2017, 01:00</td>
<td>AS</td>
</tr>
<tr>
<td>14 February 15, 2017, 06:00</td>
<td>February 16, 2017, 21:00</td>
<td>MC</td>
</tr>
<tr>
<td>15 February 21, 2017, 06:00</td>
<td>February 21, 2017, 11:00</td>
<td>MC</td>
</tr>
<tr>
<td>16 February 25, 2017, 09:00</td>
<td>February 26, 2017, 19:00</td>
<td>MC</td>
</tr>
<tr>
<td>17 March 2, 2017, 03:00</td>
<td>March 5, 2017, 07:42</td>
<td>MC</td>
</tr>
</tbody>
</table>

*Note. The duration of cold fronts and their types is identified. AS = arctic surge; MC = migrating cyclone.*

### 2.3.3 Statistical Analysis

Subtidal variations of water levels, currents, and wind stress were determined by applying a sixth order Butterworth low-pass filter with cutoff frequency of 0.6 cycles per day (40 hr). Tides in Louisiana coasts are mainly diurnal, and the period of inertial oscillation defined as $2\pi/f$ (where $f$ is the Coriolis parameter) is 24.6 hr for latitude of Barataria Bay (29.4°N), while cold fronts affect the region every 3 to 8 days (Chuang & Wiseman, 1983; Stone & Wang, 1999). The commonly used 33-hr low-pass filter was tested first, but the resulting time series showed some visually observable diurnal oscillations. Therefore, the 40-hr filter was used which effectively removed energetic fluctuations caused by tides and inertial oscillations. Water level time series were demeaned by subtracting the data mean from each observation so that the mean is zero before applying the low-pass filter. The wind stress ($\tau$), which is an indicator of energy imparted on the water surface, was calculated from the quadratic law:

$$\tau = \rho C_d |\vec{U}| \vec{U}$$

(Eq. 2.1)

where $\rho$ is the density of air (1.3 kg m$^{-3}$), $\vec{U}$ is the wind velocity at 10 m above the sea surface, and $C_d$ is the drag coefficient. The empirical formula proposed by Wu (1980 and 1994) was used for the parametrization of the drag coefficient:
The continuous wavelet transform:

\[ C_d = \begin{cases} 
C_a & U \leq W_a \\
C_a + \frac{c_b - c_a}{W_b - W_a}(u - W_a) & W_a \leq U \leq W_b \\
C_b & U \geq W_b 
\end{cases} \]  

(Eq. 2.2)

The default values for the empirical factors were \( C_a = 1.255 \times 10^{-3}, C_b = 2.425 \times 10^{-3}, W_a = 7 \) m/s and \( W_b = 25 \) m/s. The wind stress was decomposed to along-estuary and cross-estuary components. The estuary is oriented \( \sim 340 \) °T, and estuarine axis relative to the coastline is \( \sim 90 \)° therefore along-estuary winds (\( \sim 340 \) °T) can be regarded as local winds and cross-estuary winds (\( \sim 70 \) °T) that induce coastal Ekman transport can be regarded as remote winds. Positive values of along-estuary wind correspond to northwestward wind (wind blowing up the estuary) and positive cross-estuary winds represent northeastward winds (Figure 2-1).

Correlation analysis with variable time lags was used to study coherence between two time series (Kundu & Allen, 1976; Thomson & Emery, 2014). Lags were given as the temporal shifts between time series that give maximum correlation. The orientation of the principal axes of wind forcing over the water surface may influence local and remote forcing in the estuary (Holbrook et al., 1980). Therefore, principal component analysis (PCA; Kundu & Allen, 1976; Thomson & Emery, 2014) was used to define principal axis of the wind vectors during the observation period. In order to calculate the principal axes of the wind stress vector time series, first the respective east-west (\( \tau_x \)) and north-south (\( \tau_y \)) components of the wind stress at station BB7 were demeaned by removing the respective means from each record. Then, PCA was used to find the principal axes of variance along which the variance in the wind stress fluctuations \( \tau_x = (\tau_x, \tau_y) \) is maximized. The principal angles (\( \theta_p \)) were obtained by the two roots of

\[ \tan(2\theta_p) = \frac{2\tau_x\tau_y}{\tau_x^2 - \tau_y^2} \]  

(Eq. 2.3)

The detail of PCA for single vector time series is described in Thomson and Emery (2014). Because of the inherent limitations of traditional spectral methods such as Fourier analysis, which assume that the signals are stationary, wavelet coherence analysis (Grinsted et al., 2004) was used to explore variability of the wind and its interaction with subtidal water level and currents. Wavelet analysis has been extensively used to study nonstationary processes in geophysical science (Farge, 1992; Gamage & Blumen, 1993; Lau & Weng, 1995; Morlet, 1983). Three candidate wavelets were tested and the Morlet wavelet with a dimensionless frequency of six was chosen for wavelet analysis.

**2.3.4 Selection of Mother Wavelet**

The Morlet wavelet used in this study is defined as:

\[ \psi_0(\eta) = \pi^{-1/4}e^{-\iota \omega_0 \eta}e^{-1/2\eta^2} \]  

(Eq. 2.4)

where \( \omega_0 \) is dimensionless frequency and \( \eta \) is dimensionless time. The convolution of a time series of interest \( (x_n, n = 1, \ldots, N) \) with the scaled and normalized wavelet is defined to be the continuous wavelet transform:
\[ W_n^X(s) = \sqrt{\frac{\delta t}{s}} \sum_{n'=1}^{N} x_{n'} \psi_0 \left[ (n' - n) \frac{\delta t}{s} \right] \]  
(Eq. 2.5)

where \( \delta t \) is unit time step of time series and \( s \) is wavelet scale so that \( \eta = s \cdot t \). The wavelet power is defined as \( |W_n^X(s)|^2 \) and \( W_n^X(s) \) can be interpreted as the local phase. Suppose \( W^X \) and \( W^Y \) denote the continues wavelet transform of time series \( x_n \) and \( y_n \), then their wavelet coherence is defined to be (Torrence & Webster, 1998):

\[ R_n^2(s) = \frac{|S(s^{-1}W_{n}^{xy}(s))|^2}{s(s^{-1}W_{n}^{x}(s))^2.s(s^{-1}W_{n}^{y}(s))^2} \]  
(Eq. 2.6)

where \( S \) is a smoothing operator, \( W^{xy} \) is the cross wavelet transform of time series \( x_n \) and \( y_n \) and is defined as \( W^{xy} = W^X \cdot W^Y * \), where * denotes complex conjugation. The significance level was determined using Monte Carlo method, and the cone of influence in which edge effects become important was calculated. Further details about smoothing operator design, significant level and cone of influence calculations can be found in (Grinsted et al., 2004).

The selection of mother wavelet is important for effective signal processing, since the results of wavelet transform may be affected by the mother wavelet selected. Hence different methods have been developed both from qualitative and quantitative aspects for the choice of mother wavelet. The qualitative methods rely on the similarity between the analyzed signal and the wavelet (Gao & Yan, 2011). Qualitative methods are generally difficult to use and have deficiencies because they are based on visual comparison. Thus, quantitative methods have been introduced. Here we have used Shannon Entropy and Energy-to-Shannon Entropy Ratio (ESER) to select the appropriate mother wavelet. According to the minimum Shannon entropy criterion, the mother wavelet minimizing the Shannon entropy is chosen as the most suitable wavelet and according to the maximum energy-to-Shannon entropy ratio criterion, the wavelet which gives the maximum ratio is selected as the mother wavelet (Gao & Yan, 2011). ESER is defined as:

\[ ESER (s) = \frac{E_{\text{energy}}(s)}{E_{\text{entropy}}(s)} \]  
(Eq. 2.7)

where \( E_{\text{energy}}(s) \) and \( E_{\text{entropy}}(s) \) are the energy and Shannon entropy associated with each scaling parameter \( s \) respectively and are defined as:

\[ E_{\text{energy}}(s) = \sum_{n=1}^{N} |wt(s, n)|^2 \]  
(Eq. 2.8)

\[ E_{\text{entropy}}(s) = -\sum_{n=1}^{N} p_n \log_2 p_n \]  
(Eq. 2.9)

where \( N \) is the number of wavelet coefficients and \( wt(s, n) \) represents the wavelet coefficients, \( p_n \) is the energy probability distribution of the wavelet coefficients, defined as:

\[ p_n = \frac{|wt(s,n)|^2}{E_{\text{energy}}(s)} \]  
(Eq. 2.10)
Since this paper focuses on subtidal processes, corresponding scales for subtidal frequencies (3-7 days) are chosen for the evaluation of each wavelet. Three candidate mother wavelets were preselected. The Shannon Entropy and Energy-to-Shannon Entropy Ratio (ESER) extracted from water levels at station BB1, along-estuary current at the mouth (BB1) and wind stress time series (BB7) by each wavelet is listed in Table 2-2. According to this table, the wavelet which has a minimum Shannon Entropy and a maximum Energy-to-Shannon Entropy Ratio is Morlet for all three signals. It therefore represents the best-suited mother wavelet.

Table 2-2 Shannon Entropy and Energy-to-Shannon Entropy Ratio for Different Mother Wavelets and Signals

<table>
<thead>
<tr>
<th>Signal</th>
<th>Wavelet</th>
<th>Shannon Entropy</th>
<th>Energy-to-Shannon Entropy Ratio (ESER)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water level</td>
<td>Morlet</td>
<td>15.01</td>
<td>156.67</td>
</tr>
<tr>
<td></td>
<td>Paul</td>
<td>16.13</td>
<td>153.57</td>
</tr>
<tr>
<td></td>
<td>Dug</td>
<td>22.41</td>
<td>148.32</td>
</tr>
<tr>
<td>Along-estuary current</td>
<td>Morlet</td>
<td>28.36</td>
<td>150.33</td>
</tr>
<tr>
<td></td>
<td>Paul</td>
<td>32.2</td>
<td>146.92</td>
</tr>
<tr>
<td></td>
<td>Dug</td>
<td>36.20</td>
<td>142.01</td>
</tr>
<tr>
<td>Wind Stress</td>
<td>Morlet</td>
<td>20.52</td>
<td>152.87</td>
</tr>
<tr>
<td></td>
<td>Paul</td>
<td>21.30</td>
<td>151.67</td>
</tr>
<tr>
<td></td>
<td>Dug</td>
<td>30.02</td>
<td>149.44</td>
</tr>
</tbody>
</table>

2.3.5 Surface Slope

To further examine the relationship between wind and surface slope, the expected local wind-induced surface slope was calculated based on a force balance between surface and bottom shear stress and surface slope along the estuary. Assuming a quasi-steady state condition, along-estuary, vertically integrated momentum balance can be written as:

$$\frac{\partial \eta}{\partial x} = \frac{\tau_{al} - \tau_b}{\rho g h}$$

(Eq. 2.11)

where the advection of momentum has been neglected as well as Coriolis force. The flow is assumed to be barotropic so that the density gradient was omitted. This steady momentum balance has been previously applied in estuaries (e.g., Huang & Li, 2017; Janzen & Wong, 2002) and continental shelves (e.g., Csanady, 1982). In equation 2.11, $\rho$ is the density of the water (1,027 kg/m³), $h$ is the mean depth of the bay (2 m), $g$ is the acceleration due to gravity (9.8 m/s²), $\tau_{al}$ is the wind stress component along the estuary, and $\tau_b$ is the bottom shear stress defined by:

$$\bar{\tau}_b = \rho C_b |\bar{v}| \bar{v}$$

(Eq. 2.12)

where $\bar{v}$ is the current velocity, assumed to be 0.1 m·s⁻¹ based on observations by the acoustic current meters (Figure 2-8), and $C_b$ is the bottom drag coefficient with a conventional value of 5.0×10⁻³ (Li, 2003).
Observed water level slope was calculated as:

$$\frac{\Delta \eta}{\Delta x} = \frac{\eta_{BB1} - \eta_{BB5}}{x_{BB1} - x_{BB5}}$$  \hspace{1cm}  \text{(Eq. 2.13)}$$

where \(x_{BB1} - x_{BB5} (~23 \text{ km})\), is the along estuary distance between stations BB1 and BB5 and \(\eta_{BB1}\) and \(\eta_{BB5}\) are the water levels at the bay mouth and head, respectively.

### 2.3.6 Analytical Model

An analytical model was applied to examine the role of remote and local forcing in governing water level variability of the bay interior, as well as the surface slope and current velocity at the estuary mouth. Garvine (1985) developed a simple model of estuarine subtidal fluctuations forced by local and remote forces. Feng and Li (2010) modified Garvine's model, to allow a clockwise rotating cold front wind. The full details of model development and equation derivation are described in Garvine (1985) and Feng and Li (2010). Subtidal water elevations are described by:

$$\eta(x, t) = \text{Re}\left\{ \frac{i}{K \cosh \left( \frac{K_0}{c} \right)} \left[ e^{\frac{\alpha \pi T}{|\rho|}} \cosh \left( -\frac{K_0 (l - x)}{c} \right) - e^{i \theta} \right] \right\}$$  \hspace{1cm}  \text{(Eq. 2.14)}$$

where \(\alpha\) is a remote wind coefficient which relates the subtidal sea level at the estuarine mouth to cross-shelf Ekman flu, \(\omega\) is angular velocity of wind \((2\pi T\) where \(T\) is the cold front period), \(\tau\) is the wind stress, \(\theta\) is initial phase, \(l\) is the length of the estuary, \(\text{e} = (gh)^{\frac{1}{2}}\) is the linear shallow water wave speed, \(f\) is the Coriolis parameter, \(\rho\) is the water density, and \(K\) is a complex wave number of order unity given by:

$$K \equiv (-1 + i\lambda)^{\frac{1}{2}} = \left[ \frac{r-1}{2} \right]^{\frac{1}{2}} + i \left[ \frac{r+1}{2} \right]^{\frac{1}{2}}$$  \hspace{1cm}  \text{(Eq. 2.15)}$$

where \(r \equiv (1 + \lambda^2)^{\frac{1}{2}}\). \(\lambda \equiv \frac{C_b u_r}{\rho \omega}\) is a dimensionless parameter, \(C_b\) is the bottom drag coefficient, \(u_r\) is the root mean square subtidal current, and \(h\) is the mean water depth in the bay. One of the advantages of the analytical solution (Equation 2.14) is that it represents local and remote induced subtidal water levels separately. In equation 2.14, the \(\cosh\) function which is proportional to \(\alpha\), represents the water level distribution induced by remote wind and the \(\sinh\) term denotes water level distribution induced by local winds. These terms are represented separately by equations 2.16 and 12.17:

$$\eta_{\text{Remote}}^{(x,t)} = \text{Re}\left\{ \frac{i}{K \cosh \left( \frac{K_0}{c} \right)} \left[ e^{\frac{\alpha \pi T}{|\rho|}} \cosh \left( -\frac{K_0 (l - x)}{c} \right) \right] e^{i \theta} \right\}$$  \hspace{1cm}  \text{(Eq. 2.16)}$$

$$\eta_{\text{Local}}^{(x,t)} = \text{Re}\left\{ \frac{i}{K \cosh \left( \frac{K_0}{c} \right)} \left[ - e^{i \theta} \sinh \left( -\frac{K_0 (l - x)}{c} \right) \right] \right\}$$  \hspace{1cm}  \text{(Eq. 2.17)}$$

where \(\text{Re}\) implies the real part of a complex number. A major weakness of the analytical model is that it assumes zero water level change induced by local wind at the mouth of the
estuary and the boundary condition excludes the local wind-induced water levels generated outside of the estuary that can propagate into the estuary. To estimate water levels at the estuarine mouth, the starting point was set on the continental shelf, 6 km from the mouth of the estuary. A constant wind speed of 11 m/s was assumed, and wind stress was calculated from the quadratic law (equation 2.1). The surface drag coefficient was set to 1.45 × 10^{-3} according to the empirical formula proposed by Wu (1980, 1994.) (equation 2.2). The period of cold front-induced wind was assumed to be 3.5 days. Therefore, the angular frequency of the wind was calculated as 2\pi/T = 2.07 \times 10^{-5} \text{s}^{-1}. The Coriolis frequency \( f \) was set to 7.12 \times 10^{-5} \text{s}^{-1}, corresponding to Barataria Bay latitude of 29.3°N. A density of seawater of 1,027 kg/m\(^3\) was used. A bottom drag coefficient of 5.0 \times 10^{-3} (Li, 2003) was used, and root mean square subtidal current was estimated to be ~0.1 m/s. The remote wind coefficient \( \alpha \) is an empirical, geographically dependent coefficient that was used to relate remote wind with subtidal water levels. For example, based on observations in Chesapeake Bay and the Delaware Estuary (Wong & Garvine, 1984), the value of \( \alpha \) was estimated to be on the order of 5 \times 10^{-4} \text{m}^2\cdot\text{s}\cdot\text{kg}^{-1} (Garvine, 1985). Compared to deep Chesapeake Bay (~14 m) and the Delaware Estuary (~10 m), Louisiana estuaries are shallow (~2 m) and characterized by extensive marsh areas that increase the frictional damping effect on the remote wind-induced variations. Feng and Li (2010) estimated \( \alpha \) to be 8 \times 10^{-4} for Atchafalaya Bay in southern Louisiana.

2.4 Numerical Model

2.4.1 Model Setup

The FVCOM (Chen et al., 2003) was utilized to simulate the estuarine hydrodynamics under different wind and outer-shelf conditions. FVCOM is capable of calculating the nonsteady free-surface flows and transport processes in two (depth averaged) or three dimensions. The model solves the governing equations using the finite-volume method with sigma-stretched coordinate in the vertical direction, which has been used successfully in several studies of the northern Gulf of Mexico (Wang & Justic, 2009; Huang et al., 2011; Li, White, et al., 2011; Huang & Li, 2017). The system of governing equations consists of the momentum, continuity, temperature, salinity and density equations under the Boussinesq approximation and hydrostatic assumption. The detailed formulations and numerical aspects of FVCOM have been presented in Chen et al. (2003) and they will not be repeated here. The use of a barotropic 2-D model is justified considering that Barataria Bay is very shallow (~2 m) and that the water column was well mixed during the cold front season. In addition, the model is only used to simulate water levels, which is mostly a barotropic process. The numerical model mesh for this study is shown in Figure 2-2. The computational domain consists of 72,324 cells and 35,052 nodes. The grid resolution varies from about 3,000 m on the continental shelf to 15 m in channels within the estuary. Bathymetric data were obtained from Digital Elevation Models (DEM) developed by NOAA’s National Geophysical Data Center (NGDC) with 10 m resolution and were interpolated to the grid. The model domain extends from the continental shelf (40 km south of the Grand Isle) to the Lac des Allemands in the upper reaches of the estuary. FVCOM incorporates the flooding and drying treatment method to take into account the water level fluctuations over intertidal regions. The open boundary depicted as a curve between the South West Pass of the Mississippi River and Port Fourchon (southernmost part of the model domain). There were NOAA-observed water levels at both ends of the open boundary (South West Pass and Port Fourchon), thus the
linear interpolation was performed using the two water level time series to define the water levels at the open boundary.

Figure 2-2 The FVCOM model grid of Barataria Bay. NOAA = National Oceanic and Atmospheric Administration.

2.4.2 Numerical Model Experiments

A series of four numerical modeling experiments was designed that include standard model run (NME1), model validation (NME2), evaluating the effect of local winds (NME3), and evaluating the effect of remote winds (NME4; Table 2-3). Four performance indices were used to evaluate the modeling skill. The coefficient of efficiency $E$ (Nash & Sutcliffe, 1970) was calculated as:

$$E = 1 - \frac{\sum_{i=1}^{N} (OBS_i - SIM_i)^2}{\sum_{i=1}^{N} (OBS_i - \overline{OBS})^2}$$  \hspace{1cm} (Eq. 2.18)

where $OBS_i$ and $SIM_i$ are the observed and simulated values at each time step, respectively, $\overline{OBS}$ is the mean of the observed values and $N$ is the size of the data. $E$ is a measure
of goodness of fit and ranges from $-\infty$ to 1 with larger values indicating a better fit. Generally, $E > 0$ is considered an excellent model fit to the data.

Index of agreement (IOA) (Willmott, 1981) represents the correlation between observed and simulated values and was calculated according to the following equation:

$$ IOA = 1 - \frac{\sum_{i=1}^{N} (OBS_i - SIM_i)^2}{\sum_{i=1}^{N} |SIM_i - OBS| + |OBS_i - OBS|^2} \quad (Eq. \ 2.19) $$

The root mean squared error (RMSE) was calculated as:

$$ RMSE = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (OBS_i - SIM_i)^2} \quad (Eq. \ 2.20) $$

Finally, the mean absolute error (MAE) was evaluated as:

$$ MAE = \frac{1}{N} \sum_{i=1}^{N} |OBS_i - SIM_i| \quad (Eq. \ 2.21) $$

Model sensitivity analysis revealed that bed resistance and eddy viscosity to be the most sensitive model parameters. The bottom roughness height was set to 0.001 m, and the Smagorinsky coefficient was then adjusted iteratively to give the best agreement with the observed water levels at six bay stations. The Smagorinsky coefficient of 0.15 was finally adopted for the experiments NME2–NME4.

NME-3 was designed to assess the effects of local winds and cold fronts on the water levels. In this experiment, the offshore open boundary was prescribed with astronomical tides. A spatially uniform wind was applied on the surface to account for local wind effects. Local wind-induced subtidal water levels were obtained by applying a 40-hr low-pass filter.

### Table 2-3 Numerical Model Experiments

<table>
<thead>
<tr>
<th>Open Boundary Condition</th>
<th>Wind Forcing</th>
<th>Simulation Period</th>
<th>Purpose of the Experiment</th>
</tr>
</thead>
<tbody>
<tr>
<td>NME-1</td>
<td>Astronomical tides + meteorological tides (Observed)</td>
<td>Yes</td>
<td>December 7, 2015 - January 20, 2016</td>
</tr>
<tr>
<td>NME-2</td>
<td>Astronomical tides + meteorological tides (Observed)</td>
<td>Yes</td>
<td>January 20, 2016 - March 5, 2016</td>
</tr>
<tr>
<td>NME-3</td>
<td>Tides</td>
<td>Yes</td>
<td>December 7, 2015 - March 5, 2016</td>
</tr>
<tr>
<td>NME-4</td>
<td>Astronomical tides + meteorological tides (Observed)</td>
<td>No</td>
<td>December 7, 2015 - March 5, 2016</td>
</tr>
</tbody>
</table>

NME-4 was designed to examine the effects of remote force. This experiment used observed water level (astronomical tides plus meteorological tides) at the offshore boundary. No
wind was applied to the model. Remote wind-induced subtidal water levels were then estimated by applying a 40-hr low-pass filter.

2.5 Results

2.5.1 Statistical Analysis Results

2.5.1.1 PCA of Winds

The orientation of the principal axes of wind over a water surface may influence the relative contribution of local and remote forcing in an estuary. To quantify seasonal changes in wind forcing, principal wind stress during summer (21 July 2016 to 22 October 2016) was compared with the wind stress during the cold front season (7 December 2016 to 5 March 2017; Figure 2-3). PCA showed that during the summer the principal axis of wind stress was from northeast (37°T) and accounted for 60% of the variability in the wind stress (Figure 2-3a).

![Figure 2-3 Variability of wind stress in Barataria Bay for the two principal components during (a) summer and (b) cold front season. Each red line indicates the starting point of a single cold front, that is, when southerly winds change to northerly. The angle of the principal wind stress and the variance explained are shown at the top of the panels.](image)

During the cold front season, the principal axis was primarily from north-northwest (350°T) which corresponds to the main axis of the estuary (340°T; Figure 2-3b). The variance for the summer was low compared to the cold front season (Figure 2-3). Wind roses also show that the distribution of the wind direction is different during summer compared to the cold front
season (Figure 2-4). The most prevalent wind direction tends to be northeast-southwest during the summer, while it shifts to northwest-southeast direction during the cold front season (Figure 2-4).

![Wind Speed (m.s⁻¹)](image)

Figure 2-4 Comparison of wind roses (a) during the summer and (b) during the cold front season (b). Wind directions are expressed as the directions from which the wind blows (meteorological convention).

### 2.5.1.2 Relationship Between Wind and Water Level

Cold fronts are associated with an initial setup in water levels during a prefrontal phase (southerly winds) followed by a subsequent set down during frontal phase (northerly winds). Water level change (difference between setup and subsequent set down levels) is always higher at the head (Figure 2-5e). For example, the maximum water level change during a cold front that affected the region on 22 January 2017 was 0.55 m in Bay Jimmy (BB3) and 0.37 m at the mouth (BB1). The variation in subtidal water levels is indicated by the range in standard deviations which increases from 0.08 m at the mouth to 0.14 m at the bay head (Table 2-4). Both highest and lowest subtidal water levels are found near the head where along-estuary wind effect is the strongest. Water level variability is influenced by the east to west orientation of the Louisiana coastline (Feng & Li, 2010; Perez et al., 2000).
Figure 2-5 The time series of along-estuary wind stress ($\tau_{al}$) and subtidal water levels at (a) BB1, (b) BB2, (c) BB3, and (d) BB5. (e) A comparison of water levels at stations BB1, BB2, BB3, and BB5 and cross-estuary wind stress ($\tau_{cr}$).

Table 2-4 Correlation Coefficients ($R$) and Lag Times for Along-Estuary/Cross-Estuary Winds and Water Levels at BB1 ($\eta_{BB1}$), BB2 ($\eta_{BB2}$), BB3 ($\eta_{BB3}$), BB5 ($\eta_{BB5}$), Water Level Difference ($\Delta\eta$), and Along-Channel Current ($u_{al}$)

<table>
<thead>
<tr>
<th></th>
<th>Along-estuary wind stress</th>
<th>Cross-estuary wind stress</th>
<th>Standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R</td>
<td>Lag time (h)</td>
<td>R</td>
</tr>
<tr>
<td>$\eta_{BB1}$</td>
<td>0.24</td>
<td>20</td>
<td>0.45</td>
</tr>
<tr>
<td>$\eta_{BB2}$</td>
<td>0.41</td>
<td>3</td>
<td>0.30</td>
</tr>
<tr>
<td>$\eta_{BB3}$</td>
<td>0.62</td>
<td>2</td>
<td>0.42</td>
</tr>
<tr>
<td>$\eta_{BB5}$</td>
<td>0.5</td>
<td>3.5</td>
<td>0.38</td>
</tr>
<tr>
<td>$\Delta\eta$</td>
<td>0.80</td>
<td>5</td>
<td>0.20</td>
</tr>
<tr>
<td>$u_{al}$</td>
<td>0.14</td>
<td>14</td>
<td>0.75</td>
</tr>
</tbody>
</table>

Water levels at the bay head ($\eta_{BB5}$) and Bay Jimmy ($\eta_{BB3}$) appear well correlated and nearly in phase with along-estuary wind stress ($\tau_{al}$; Figures 2-5c and 2-5d). The time lag between
water levels and along-estuary wind stress at stations BB3 and BB5 which makes maximum correlation coefficient, are 2 and 3.5 hr, respectively (Table 2-4). Correlation coefficients for $\tau_{al}$ and water levels at stations BB1, BB2, and BB3 were 0.24, 0.41, and 0.62, respectively. However, correlation coefficient drops to 0.5 at station BB5, which suggests that other factors in addition to local wind influence water levels at this location (Table 2-4).

![Wavelet coherence for along-estuary wind stress ($\tau_{al}$) and water levels](image)

Figure 2-6 Wavelet coherence for along-estuary wind stress ($\tau_{al}$) and water levels at (a) BB1 ($\eta_{BB1}$), (b) BB2 ($\eta_{BB2}$), (c) BB3 ($\eta_{BB3}$), and (d) BB5 ($\eta_{BB5}$). The 5% significance level against red noise is shown as a thick contour. The U-shaped curves with lighter shade of the power spectrum show the cone of influence where edge effects become important.

To illustrate the coherency in time frequency domain, wavelet coherence between $\tau_{al}$ and each set of water level records was computed (Figure 2-6). Arrows in Figure 2-6 are more or less pointing to the right (0°) which indicates largely in phase relation between water level and along-estuary wind stress in all stations except BB1 (see arrows in Figure 2-6). $\tau_{al}$ was most coherent with water levels around 3- to 10-day cycle period. Coherency between $\tau_{al}$ and water levels increases from the bay mouth (BB1) to the bay head (BB5). This suggests that local wind effect is the dominant controlling mechanism at the head, and in contrast, remote wind effect is dominant at the mouth. When remote wind blows parallel to the shoreline, a dynamic balance is established between the wind-induced frictional force and the Coriolis force. Therefore, a mass
transport of water takes place in the surface layer, which is directed to the right of the wind. Depending on the remote wind direction relative to the coastline, water level is lowered or elevated along the shoreline. The resulting disturbances then propagate into the estuary and contribute to the water level variations.

### 2.5.1.3 Wind and Surface Slope Relationships

The maximum water level difference between the mouth (BB1) and stations BB2, BB3, and BB5 is 0.24, 0.27, 0.28, and 0.30 m respectively (Figure 2-5e), which is relatively significant compared to the tidal range (maximum tidal range is ~0.62 m). To further examine the relationship between wind and surface slope, the calculated local wind-induced surface slope and observed surface slope were calculated based on equation 2.11 and 2.13, respectively. Figure 2-7 shows the expected and observed water level slope. Both slopes are of the same order of magnitude and well correlated ($R = 0.87$). Although a constant depth is being used in the steady state solution (equation 2.11), it predicts the observed water level slope well. This result indicates that local wind is the dominant forcing in determining the water level slope inside Barataria Basin.

![Figure 2-7 Along-estuary surface slopes between the bay head (BB5) and bay mouth (BB1); observed (solid blue line) and estimated from along-estuary wind stress (dashed green line). Positive values indicate $\eta_{BB1} > \eta_{BB5}$. Vertical dashed lines denote cold fronts passing through Barataria Bay.](image)

### 2.5.1.4 Relationships Between Wind, Water Level, and Bay Mouth Currents

Along-estuary current at the bay mouth ($u_{al}$), water level at the bay mouth ($\eta_{BB1}$), water level difference ($\Delta \eta$) between stations BB5 and BB1, along-estuary wind stress ($\tau_{al}$), and cross-estuary wind stress ($\tau_{cr}$) are compared next to characterize relationships between the current at the bay mouth and different forcing factors (Figure 2-8). As shown in section 4.1.3, the interior bay slope is induced by local wind forcing and not by the remote winds, whereas $\eta_{BB1}$ is forced largely by remote winds. Therefore $\Delta \eta$ and $\tau_{al}$ are used as proxies for the local forcing, while $\eta_{BB1}$ and $\tau_{cr}$ are used as proxies for the remote forcing. Most notable are the differences in the patterns of variability in $\Delta \eta$ and $\tau_{al}$ (local effect) and measured $u_{al}$ (Figures 2-8a and 2-8c), which indicates no clear relationship between local forcing and bay mouth currents except during cold fronts when strong outflow currents are observed. In contrast, $\tau_{cr}$ and $u_{al}$ are clearly correlated (Table 2-4) and in phase such that a negative cross-estuary (westward) wind stress corresponds
to a negative up-estuary current (Figure 2-8d). This is not surprising given that the remote forcing is expected to be the strongest at the bay mouth (section 4.1.2).

Wavelet coherence analysis was used to elucidate relationships between \( u_{AL} \) and different forcing factors including \( \Delta \eta \), \( \tau_{AL} \), \( \eta_{BB1} \), and \( \tau_{CR} \). The remote forcing is significantly coherent with \( u_{AL} \) at low frequencies (Figures 2-9b and 2-9d). In contrast, the local forcing is not significantly coherent with \( u_{AL} \) (Figures 2-9a and 2-9c). These results suggest that remote wind has the controlling influence on \( u_{AL} \).
Figure 2-9 Wavelet coherence between along-estuary current at the bay mouth $u_{AL}$ and (a) along-estuary water level difference ($\Delta \eta$), (b) subtidal water level at the bay mouth ($\eta_{BB1}$), (c) along-estuary wind stress ($\tau_{AL}$), and (d) cross-estuary wind stress ($\tau_{CR}$). The 5% significance level against red noise is shown as a thick contour. The U-shaped curves with lighter shade of the power spectrum show the cone of influence where edge effects become important.

2.5.2 Analytical Model Results

Analytical model was run from 21 January 2017 to 25 January 2017. A single cold front passed the region during this period (Figure 2-10a). The idealized cold front wind simulated by the analytical model is shown in Figure 2-10b. Results were compared with observed subtidal water levels at six stations within Barataria Bay for different values of $\alpha$ (Figure 2-10). Note that $\alpha$ is the remote wind coefficient that relates the subtidal sea level at the estuarine mouth to cross-shelf Ekman flux. A value of $\alpha = 4 \times 10^{-5}$ m$^2$.s$^{-1}$.kg$^{-1}$ appears to best describe the observed subtidal water levels. Using the equations 2.16 and 2.17, we could describe the relative importance of local and remote winds in governing subtidal water level variations for the entire estuary. As shown in Figure 2-11b, the analytical model reveals a greater local wind-induced amplitude at the bay head, with a slope toward the mouth where the water level variation is minimal. Under idealized cold front winds, the maximum water level amplitude for the bay head and the bay mouth was 0.5 and 0.1 m, respectively. The remote wind-induced water levels are
almost constant along the estuarine longitudinal axis and decrease only slightly with distance from mouth (Figure 2-11a). This is because the length of the estuary is small compared to low-frequency wavelength, and so the remote wind-induced water level is basically uniform along the bay, a pattern referred to as Helmholtz mode, piston mode, or pumping (Honda et al., 1908). Consequently, the primary factor affecting water level slope is the local wind forcing (Figure 2-11b). Results of analytical model suggest that the effects of local winds on water level variability are important and can account for ~60% of the total subtidal water level variability at the bay head and ~50% of the variability at the bay mouth.

Figure 2-10 Wind speed observed at station BB7; (b) idealized clockwise rotating wind input to the analytical model; (c) measured and predicted subtidal water levels at BB1; (d) measured and predicted subtidal water levels at BB2. The black lines denote observed subtidal water levels and the yellow lines denote predicted subtidal water levels for $\alpha = 4 \times 10^{-5}$.
2.5.3 Numerical Model Results

The simulated water levels are in good agreement with the observed water levels (Figure 2-12; Table 2-5). For example, the average values of the coefficient of efficiency and the index of agreement are 0.25 and 0.90, respectively, indicating that simulated and observed water levels are highly correlated. The mean absolute error at the bay head (station BB6) is 0.06 m and increases to 0.1 m at the bay mouth (station BB1), suggesting that the model is sufficiently accurate for evaluating the effect of the local and remote forcings.
Figure 2-12 Measured and simulated water levels at the six stations in Barataria Bay (validated model results NME-2)

Table 2-5 Performance Indices Between the Validated Model (NME-2) and Observed Data at Each Station

<table>
<thead>
<tr>
<th></th>
<th>BB1</th>
<th>BB2</th>
<th>BB3</th>
<th>BB4</th>
<th>BB5</th>
<th>BB6</th>
</tr>
</thead>
<tbody>
<tr>
<td>$E$</td>
<td>-0.21</td>
<td>0.05</td>
<td>0.10</td>
<td>0.35</td>
<td>0.60</td>
<td>0.65</td>
</tr>
<tr>
<td>$IOA$</td>
<td>0.79</td>
<td>0.89</td>
<td>0.92</td>
<td>0.95</td>
<td>0.94</td>
<td>0.96</td>
</tr>
<tr>
<td>$RMSE$</td>
<td>0.12</td>
<td>0.11</td>
<td>0.10</td>
<td>0.08</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>$MAE$</td>
<td>0.10</td>
<td>0.09</td>
<td>0.08</td>
<td>0.08</td>
<td>0.07</td>
<td>0.06</td>
</tr>
</tbody>
</table>

*Note. $E$ = coefficient of efficiency; $IOA$ = index of agreement; $RMSE$ = root mean square error; $MAE$ = mean absolute error.*

2.5.3.1 Effects of Local and Remote Winds

Numerical model results indicate that during cold fronts, Barataria Bay exhibits larger setup/set down at the bay head compared to the bay mouth (Figure 2-13a). This is consistent with the observations and analytical model results. The standard deviation of local wind-induced subtidal water levels at the bay head and the bay mouth was 0.09 and 0.16 m, respectively. Generally, water level set down associated with frontal phase was larger than water level setup in prefrontal phases. Local winds contribute to a significant fraction of the total water level variations. For example, the observed subtidal water level on 8 January 2017 was 0.5 m and local wind-induced water level was 0.32 m in station BB4. Therefore, local wind effects could account for >50% of the total water level variation.
Figure 2-13 Subtidal water levels predicted by the numerical model for six stations within Barataria Bay: (a) locally induced (NME 3) and (b) remotely induced (NME 4). Water level difference: (c) induced by local winds (NME 3) and (d) induced by remote winds (NME 4). Vertical black lines denote cold fronts passing through Barataria Bay.

Local wind-induced water level differences between the head and the mouth of the bay show significant variations, ranging from −0.25 to 0.16 m (Figure 2-13c). Negative water level differences (blue areas in Figure 2-13c) correspond to frontal phases and positive water level differences (red areas in Figure 2-13c) are indicative of prefrontal phases. Results of NME3 experiment suggest that local wind contributes significantly to both the total water level variation and water surface slope. The amplitude of the remote wind-induced subtidal water levels (Figure 2-13b) is generally smaller to the one induced by local winds (Figure 2-13a). Likewise, the scale of the remote wind-induced differences in water levels (Figure 2-13d) is smaller than that of local winds. The standard deviation of the local wind-induced water level differences (0.08 m) was 8 times larger than for the remote wind-induced water level differences (0.01 m). This affirms that local wind strongly affects water level slope whereas the effects of remote wind are negligible (Figures 2-13c and 2-13d).
2.6 Discussion

During the 2016–2017 cold front seasons, the principal wind axis was generally oriented along the axis of the estuary which further exacerbated the role of the local wind forcing. Wavelet coherence analysis revealed that water levels inside the bay are coherent with along-estuary wind stress over 3- to 10-day periods, consistent with the passage of cold fronts over the northcentral Gulf of Mexico (Marmorino, 1982; Moeller et al., 1993; Roberts et al., 1989). Barataria Bay exhibits larger setup/set down during southeasterly/northwesterly at the bay head compared to the bay mouth. This is because the estuary becomes progressively shallower at the head which favors direct setup caused by local winds. Further, multiple coherent analysis has shown that remote wind was more coherent with subtidal currents at the bay mouth compared to local wind, suggesting that estuarine-shelf exchanges are facilitated by coastal Ekman transport. However, local winds can also induce strong outflows during cold fronts.

Both analytical and numerical model results indicate that the remote wind-induced water levels are spatially uniform within the bay (i.e., Helmholtz mode). This condition is largely due to the geomorphology of Barataria Bay, that is, relatively narrow inlets that connect the estuary to the Gulf of Mexico shelf and relatively small length of the bay compared to the scale of the remote wind-induced waves. The Helmholtz mode has been widely employed to describe the tidal response of narrow-mouthed bays and harbors (Maas, 1997; Miles & Lee, 1975; Molines et al., 1989) and large systems such as Gulf of Mexico to explain the relatively large amplitudes of diurnal tides (Platzman, 1972).

Despite remote wind effects contributed significantly to the subtidal water levels variability, they were not the dominant component in Barataria Bay. Indeed, the magnitude of water level variations associated with local forcings was at least twice as large as those associated with remote forcings, especially at the bay head. Our results also differ from other studies in the region, notably in Breton Sound (Snedden et al., 2007) and Lake Pontchartrain (Huang & Li, 2017), where remote wind dominates over local winds. We suggest that these differences are mainly due to the differences in geomorphology of estuaries. Barataria Bay is located to the west of the Mississippi River where the adjacent shelf and alongshore currents are intercepted by the Mississippi delta. Barataria Bay is also constricted on the southern end by a chain of barrier islands where the bay-shelf exchanges are only possible through narrow inlets. Therefore, wind-driven cross-shelf Ekman transport is not fully developed, which further reduces water level co-oscillations between the shelf and the estuary. Barataria Bay is almost perpendicular to the coast (340°T). Thus, local and remote forces operate nearly independently, since the remote effect is proportional to the alongshore component while the local effect is proportional only to the onshore component. In contrast, in Lake Pontchartrain, the two wind mechanisms act in concert with each other since estuarine axis is nearly parallel to the coast. The orientation of the estuary relative to the prevalent direction of local winds also has a controlling effect on the degree of local wind dominance. For Barataria Bay, the local effect is expected to be greater since the bay’s orientation (340°T) corresponds to the prevalent direction of local winds. This does not apply to Lake Pontchartrain.

Another reason for the dominance of local wind effect in Barataria Bay is its relatively shallow depth (~2 m) compared to Lake Pontchartrain (~4–5 m) which favors water level setup/set down in response to local winds. We now use the analytical model (Feng & Li, 2010; Garvine, 1985) to further examine the effect of estuarine depth on local and remote wind-induced water level variations. Estuarine depths of 2, 4, and 6 m were used, and their respective

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water level variations at station BB3 were calculated. Results show that remote wind-induced water level variations are almost insensitive to estuarine depth (Figure 2-14a). In contrast, local wind-induced water levels are highly sensitive to estuarine depth (Figure 2-14b). Amplitudes of local wind-induced water level fluctuations range from 0.31, 0.26, and 0.23 m for estuarine depths of 2, 4, and 6 m respectively. Thus, the depth of an estuary is a key factor that affects the relative dominance of local or remote winds.

![Graphs of water levels](image)

Figure 2-14 Subtidal water levels predicted by the analytical model at station BB3 for three different estuarine depths within Barataria Bay: (a) remotely induced, (b) locally induced, and (c) total.

Breton Sound estuary is also shallow (~1 m) but its highly complex hydrologic features that include vegetated marshes, tidal flats, and abandoned delta distributaries frictionally dissipate waves propagating through the system (Snedden et al., 2007). Consequently, contrary to Barataria Bay where remote wind-induced fluctuations exhibit little or no frictional attenuation, significant damping of remote wind-induced subtidal fluctuations occurs in Breton Sound estuary as a result of frictional damping (Snedden et al., 2007).

### 2.7 Conclusion

The analysis of subtidal water levels and bay mouth current fluctuations in Barataria Bay provides new insight into the dynamical structure of complex bar-built estuaries. Barataria Bay exhibits water level variations that are largely affected by the passage of cold fronts over the northcentral Gulf of Mexico. The local and remote wind forcings are of about equal importance
in controlling the overall water level variations at the bay mouth, whereas the local forcing is the dominant driver inside the bay. On the other hand, remote wind forcing dominates the subtidal currents at the bay mouth and thus controlling the exchange between the estuary and the ocean. However, local winds can also induce strong outflows during cold fronts.

The results of this study underscore the importance of local and remote winds in governing estuarine water level and current dynamics. Understanding estuarine water level and current variations is instrumental for assessing the effects of hydrologic restoration, river flow management, and the ecological and socioeconomic consequences associated with prolonged inundation. For example, Barataria Bay is the site of major coastal restoration efforts, where two large-scale river diversions are currently under consideration (Coastal Protection and Restoration Authority, 2017). Those river diversions will likely strongly affect the subtidal water levels in the vicinity of the diversion sites that could lead to flooding of nearby coastal communities and negatively impact marshes by increasing inundation. In this regard, understanding the effects of local and remote forcing on estuarine water levels greatly enhances our ability to accurately predict the consequences of proposed river diversions.
CHAPTER 3. SUSPENDED SEDIMENT DYNAMICS IN A DELTAIC ESTUARY CONTROLLED BY SUBTIDAL MOTION AND OFFSHORE RIVER PLUMES

3.1 Introduction

Connectivity between an estuary and the coastal ocean is a key factor impacting critical estuarine processes, such as nutrient exchange, regulation of estuarine residence time, acidification and hypoxia (e.g., Davis et al., 2014; O'Callaghan et al., 2007). Connection with the ocean in bar-built estuaries commonly found along the southeast and Gulf coasts, differs from that of coastal plain, tectonic, or fjord estuaries. Indeed, bar-built estuaries are typically separated from the ocean by barrier islands, and thus their exchange is concentrated through tidal inlets. Additionally, coupled interaction between an estuary and offshore freshwater and sediment sources (i.e., neighboring river) can also impact sediment dynamics (e.g., McConnaughey et al., 1994; Yamada & Kosro, 2010). The main objective of this study is to investigate the factors affecting variability in suspended sediment concentration (SSC) within the main inlet of Barataria Bay, a bar-built estuary within the Mississippi Delta (Figure 3-1), including the offshore influence of river plumes from the adjacent Mississippi River. This is particularly relevant since Barataria Bay has experienced dramatic wetland loss in the last century (Dahl, 2000).

Barataria Bay has several tidal inlets connecting with the coastal ocean of Louisiana Bight. Barataria Pass with a width of about 600 m is the main inlet of Barataria Bay. Marmer (1948) measured tidal currents in different inlets of the Barataria Bay over a 24 day period, and found that the amount of water going through Barataria Pass was about 66% of the total. Li et al., (2011) conducted an observational study in Barataria Pass to find fluxes of water and suspended sediment during a full tidal cycle. They indicated that the net flux of total suspended sediment is 8800 t and is directed out of the bay. Barataria Bay is under the influence of an anticyclonic (clockwise) gyre in the Louisiana Bight that carries much of the Mississippi River plume northward towards the Barataria Bay shoreline (Rouse and Coleman, 1976; Wiseman et al., 1976; Walker et al., 2005). Southwest Pass, located 60 km to the southeast of Barataria Pass, is the main outlet of the Mississippi River exporting 45% to 67% of the Mississippi River discharge (Etter et al., 2004; Allison et al., 2012; Li et al., 2011). This Pass is a narrow outlet (~700 m) which results in supercritical flow around the mouth of the pass with velocities greater than 2 m s⁻¹ (Sorourian et al., 2020). Freshwater discharged through the West Bay Diversion and various nearby crevasses along the western levees of the Mississippi River birds-foot delta all debouche onto the shelf. Under the influence of buoyancy and Coriolis forces, a portion of the freshwater forms the anticyclonic gyre towards Louisiana’s barrier islands while the other flows westward along the barrier islands, composed of part of the stratified Louisiana Coastal Current (Wiseman and Kelly, 1994; Walker, 1996). Subsequently, Mississippi plume waters from the shelf may penetrate into Barataria Bay through its main tidal inlet (Barataria Pass). Contrary to what Coriolis effect would produce, Li et al. (2009) indicated that the average salinity is higher on the western end than on the eastern end of the Barataria Pass. They further suggested that the

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observed difference arises from the influence of the river water coming out of the Mississippi River through the Southwest Pass of the Birdfoot Delta (Li et al., 2009, 2011).

In particular, this study focuses on the temporal changes in SSC during the winter season characterized by low river discharge and high cold front activity as well as during the spring period which typically has high river discharge and low cold front activity. During the winter months, the dominant synoptic meteorological feature in this region is the cold fronts, which typically pass through southern Louisiana with a recurrence interval of 3-8 days (Roberts et al., 1989; Moeller et al., 1993). Associated with these frontal passages is a clockwise wind field which shift from southerly during the pre-frontal phase, to westerly, and then northerly during the frontal phase. In bar-built estuaries, wind-induced low frequency variations could be of comparable or greater magnitude than astronomical tides (Conner et al., 1987; Payandeh et al., 2019). These dynamics are especially relevant for the northern Gulf of Mexico, where astronomical tidal ranges are small but strong cold fronts are the dominant synoptic weather pattern spatially during the winter season. For bay hydrodynamics, local wind forcing caused by cold fronts is the dominant force that controls water levels and water level slope within Barataria Bay (Payandeh et al., 2019). Therefore, in addition to river plume effects, the influence of cold fronts on sediment concentration is also important and merits further investigation.

![Study area and locations of monitoring stations. BB1 denotes the station where the ADV was mounted. BB1, BB2, BB4 denote stations where the salinity data were collected by the USGS. BB3 denotes the location of the USGS station where the wind measurements were carried out. Black circles denote the position of selected sampling stations along the Barataria transect (Turner et al., 2019); the first transect station is located approximately one km offshore. The positive direction of axial velocity is represented by the blue arrow.](image-url)
3.2 Methods

3.2.1 Data Acquisition

A Nortek Acoustic Doppler Velocimeter (ADV) was deployed in Barataria Pass (station BB2, Figure 3-1). The winter deployment was carried out between 8 December 2016 and 5 February 2017 and the spring deployment was made between 15 February and 16 April 2018. The deployment site, located near the east end of the Barataria Pass cross-section, had a water depth of ~3 m and the ADV was placed at 0.4 m above the bottom. Water levels and currents were burst sampled at 4 Hz for 5 min every hour. The measured pressure data were processed to remove the inverted barometric effect before converting to the water level. An Optical Backscatter Sensor (OBS) was attached to the ADV to measure turbidity simultaneously with other measurements. Water samples (1 liter) were collected using three ISCO water samplers at the same location and water depth as for turbidity measurements and a total number of 107 samples were collected. Water samples were placed in polyethylene bottles pre-washed with a 10% HCl solution. The water samples were filtered on preweighed GF/F filters (pore size = 0.7 µm) to extract suspended sediments. The filters were dried at 60 °C for 24-48 hours, or until completely dry and then weighed to determine the total mass of suspended material. The dried filters were combusted at 550 °C for 4 hours to determine total organic matter content (LOI) (Dean 1972). The sediment concentration was then correlated to the turbidity values obtained by the OBS sensor assuming a linear response of the instrument with sediment concentrations (e.g., Voulgaris and Meyers, 2004). The relation between turbidity and measured sediment concentrations (Figure 3-2) showed increased scatter at higher turbidity values and had a correlation coefficient of 0.88.

![Figure 3-2](image)

Figure 3-2 Relationship between measured SSC and calibrated turbidity from OBS sensor.

The wind data were obtained from a USGS station located in the central section of Barataria Bay (station BB3, Figure 3-1). Wind data from this station were shown to adequately represent wind stress within the interior of the estuary (Mariotti et al., 2018). The 3-h surface analyses maps produced by NOAA’s Hydrometeorological Prediction Center
were used to identify cold fronts, and meteorological measurements from USGS (station BB3, Figure 3-1) were used to pinpoint the timing of the frontal events. A total of 17 cold fronts were identified during the observation periods, of which 11 occurred during the winter and 6 during the spring.

Salinity measurements during both winter and spring deployments were obtained from three USGS stations located in Barataria Pass (BB1), north of Grand Terre Island (BB2) and in Huckelberry Bay (BB4, see Figure 3-1). Long term salinity and TSS measurements conducted by Turner et al. (2019) were used to study the temporal responses of the bay to variations in the Mississippi River discharge. These data were collected monthly in the Barataria Basin from 1994 to 2016 at 37 stations along a 129 km transect from a starting point 1 km offshore to the upper reaches of the basin. For this study, only the first 15 stations between Barataria Pass and Little Lake were used (BT1:BT15, Figure 3-1). Daily-averaged Mississippi River discharge data from 2008 to 2018 were obtained from a USGS station located at Belle Chasse. Also, hourly river discharge data for 2016, 2017 and 2018 were obtained from the same station. Satellite images from Terra and Aqua MODIS (MODerate resolution Imaging Spectroradiometer) were obtained from the Louisiana State University Earth Scan Laboratory MODIS true color web archive (www.esl.lsu.edu/MODIS/) to track the Mississippi River sediment plume along the Louisiana shelf.

### 3.2.2 Data Analysis

Subtidal variations of water levels and axial velocities were determined by applying a sixth order Butterworth low-pass filter with a cutoff frequency of 0.6 cycles per day (40 hr). The commonly used 33-hr low-pass filter was tested first, but the resulting time series showed some visually observable diurnal oscillations. Therefore, the 40-hr filter was used which effectively removed energetic fluctuations caused by tides and inertial oscillations (Payandeh et al., 2019). Water level time series were demeaned before applying the low-pass filter. The pressure data were used to compute wave statistics after removing low frequency components (less than 0.05 Hz). Time series of axial velocity and derived SSC were used to compute continuous records of SSC flux at station BB1. Total SSC flux at unit width was computed as the product of axial velocity and SSC. To identify tidal and nontidal mechanisms of sediment flux, the total SSC flux was decomposed into tidal pumping and subtidal components. The subtidal sediment flux is driven by the tidally averaged (residual) velocity and the tidally averaged SSC and is given by

\[
F_A = \bar{U} \cdot \bar{SSC}
\]

(Eq. 3.1)

where the overbars denote tidally averaged (low-pass filtered) values. The tidal pumping flux is driven by fluctuations in tidally varying velocity and SSC and is given by

\[
F_p = U' \cdot SSC'
\]

(Eq. 3.2)

where the primes denote tidal fluctuations around the tidally averaged values. Tidal values were computed by high pass filtering of U and SSC.

The energy or variance of swell waves was defined by the area under the spectrum from 0.05 to 0.4 Hz. The significant wave height for swell \(H_s\), was then defined as \(H_s \approx 4 \times (\text{variance})^{0.5}\). The remaining part of the total spectral variance was attributed to wind waves and a
corresponding significant sea height was defined similarly. The Fast Fourier Transform (FFT) was used to calculate the frequency spectrum for each signal. Wavelet coherence analysis (Grinsted et al., 2004) was used to explore variability in SSC and its interaction with water level, axial velocity and wind stress. Wavelet analysis has been extensively used to study nonstationary processes in geophysical science (Farge, 1992; Lau & Weng, 1995; Morlet, 1983). Three candidate wavelets were tested and the Morlet wavelet with a dimensionless frequency of six was chosen for wavelet analysis (Payandeh et al., 2019).

3.3 Results

3.3.1 Forcing Conditions

Due to coupled Barataria Bay and Louisiana Bight interactions, characteristics of the Mississippi River system are critical to understanding the sediment dynamic response within Barataria Bay. Peak discharge from the Mississippi River is typically delivered in spring when the overall variability in discharge is quite large (up to $2.4 \times 10^4$ m$^3$ s$^{-1}$, Figure 3-3a). Over a 10-year period (2009-2018), the variability in spring discharge was characterized by a bimodal pattern, with the first peak in early April and the second peak in late May. Comparison between the two deployment periods shows that the system was exposed to relatively low levels of discharge with a mean values of $1.3 \times 10^4$ m$^3$ s$^{-1}$ during the winter period, while the spring deployment coincided with a large discharge event ($3.6 \times 10^4$ m$^3$ s$^{-1}$) and had a mean discharge value of $3 \times 10^4$ m$^3$ s$^{-1}$ (Figure 3-3b). During the winter, the discharge averaged around $0.9 \times 10^4$ m$^3$ s$^{-1}$ from 8 to 28 December 2016 and then increased gradually to $2 \times 10^4$ m$^3$ s$^{-1}$. During the spring, the discharge further increased from around $1.3 \times 10^4$ m$^3$ s$^{-1}$ to a maximum of around $3.6 \times 10^4$ m$^3$ s$^{-1}$ on March 12, and then stabilized at around $3 \times 10^4$ m$^3$ s$^{-1}$ until the end of the deployment.

Additional factors affecting the exchange between the bay and the shelf waters include wind forcing and its subsequent influence on estuarine surface slope and coastal water levels. The distribution of wind direction and intensity during the winter and spring deployments are shown in Figures 3-4a and 3-4a, respectively. As expected, the most frequent winds during the winter were from north, northeast and east due to frequent cold front activity in this period. In contrast, the most frequent winds during the spring were from southeast and south. During the winter, a total of 11 cold fronts passed through the study area; the winds were generally strong, with velocity frequently exceeding 10 m s$^{-1}$ (Figure 3-5a). Contrary to the winter deployment, the spring deployment only had 6 cold fronts; the winds were generally weaker and rarely exceeded 7 m s$^{-1}$ (Figure 3-6a). Moreover, spring cold fronts were weaker compared to the winter cold fronts.
Figure 3-3 Mississippi River discharge at Belle Chasse; (A) the mean daily discharge (red line) and standard deviations (gray shading) over a 10 year period (2009–2018), and maximum and minimum (“extremes”) daily discharge over the 10 year period (dashed lines). (B) the mean daily discharge from January 2016 until December 2018. The blue and green horizontal bars at the top of both panels denote the study periods; winter deployment lasted from 8 December 2016 until 5 February 2017; spring deployment lasted from 15 February 2018 until 16 April 2018 and coincided with the large discharge event ($3.3 \times 10^4 \text{ m}^3 \text{ s}^{-1}$).
3.3.2 Estuarine Responses

The data collected at station BB1 were organized as hourly time series shown in Figures 3-5 and 3-6. During the winter, daily water level variations changed between 0.16 m during the neap tides and by 0.7 m during the spring tides. As demonstrated in Payandeh et al. (2019), subtidal water level variations were mostly a result of water level set-up and subsequent set-down during frontal passages (Figure 3-5b). During the pre-frontal phase, southerly winds caused coastal sea level set-up along the coast and subsequently net inflow of Gulf water through Barataria Pass, resulting in increased water levels inside the bay. After the passage of a front, as the winds turned to northerly direction, the bay water was flushed out, resulting in decreased water levels inside the bay. As the main bay outlet to the Gulf, Barataria Pass transports large volume of water and current velocities reach high levels. In addition to the tidal exchange, Barataria Pass carries the extra water draining from the bay during the frontal passages. The highest axial velocity (0.8 m s\(^{-1}\)) was observed during the ebb tide following the passage of a cold front on 22 January 2017, as a result of strong northerly winds, while the maximum flood-directed axial velocity reached 0.6 m s\(^{-1}\) (Figure 3-5c). During winter, the spring and neap tide effects on axial velocities were less clear because the transport of water was strongly modulated by winds, especially during cold fronts. Water level variations during the spring were smaller compared to the winter (Figure 3-6b), ranging from a neap minimum of ~0.10 m to a spring maximum of ~0.58 m. The highest recorded axial ebb-directed velocity was 0.77 m s\(^{-1}\).

Cold fronts during the winter produced a series of distinct wind wave events. Higher values of SSC were found in correspondence with higher wind speeds and subsequently higher sea waves (Figures 3-5a and 3-5d). Of particular interest is the cold front that occurred on January 7, 2017, during which a wind of 5 m s\(^{-1}\) blowing from the southwest was followed by winds of up to 17 m s\(^{-1}\) blowing from the northwest, thus producing a wave event superimposed to a high axial velocity and a large sea level setdown (Figure 3-7). This event created the most favorable conditions for sediment resuspension and thus produced the highest peak in SSC during the first deployment (1.84 g L\(^{-1}\)).

Contrary to the winter deployment, only 6 cold fronts affected the Louisiana coast during the spring. Although the lack of strong and frequent cold fronts would suggest that mean SSC values would be lower, it was actually much higher during the spring (0.23 g L\(^{-1}\)) compared to the winter period (0.15 g L\(^{-1}\)). This suggested that an additional mechanism, other than resuspension by cold fronts and tidal currents, was governing the SSC dynamics during the
spring. As for the winter, the highest peak in SSC during the spring was the result of a cold front on March 20, 2018. This particular cold front was characterized by a wind of 6 m s\(^{-1}\) blowing from the southeast that was followed by winds up to 12 m s\(^{-1}\) blowing from the northwest. The winds generated significant wave height up to 0.55 m and produced the highest SSC observed during both deployments (2.0 g L\(^{-1}\)). This particular cold front was not the strongest cold front during the spring, but it was nonetheless associated with the highest SSC values, suggesting higher sediment availability for resuspension at this time.

Figure 3-5 Time series of wind speed, water level, axial velocity, wave height and SSC in Barataria Pass (station BB1) during winter deployment (from 8 December 2016 to 5 February 2017). (A) wind vectors, red lines indicate cold front events; (B) observed total and subtidal water levels; (C) observed total and subtidal axial velocities; (D) significant wave height for wind sea and swell; (E) suspended sediment concentration (SSC).
Figure 3-6 Same as in Figure 3-5 except for the spring deployment (from 15 February to 16 April 2018).
Figure 3-7 Time series of wind speed, water level, axial velocity, wave height and SSC illustrating the effects a cold front that affected the area on January 7, 2017; (A) wind vectors; (B) observed total and subtidal water levels; (C) observed total and subtidal axial velocities; (D) significant wave height for wind sea and swell; (E) suspended sediment concentration. Frontal passage occurrence is indicated by a wind shift to the northerly quadrant.

Figures 3-8a and 3-8b show the plots of the wave height as a function of the wind direction at station BB1 (only significant wave heights higher than 0.2 m are reported). As expected, during the winter the waves are produced by winds blowing from the north and north east, followed by winds blowing from the south, indicating that strong winds from these directions produce high wave events in Barataria Pass (Figure 3-8a). During the spring, waves are mostly produced by southeasterly and southerly winds (Figure 3-8b). Of more interest is the connection between wind and higher SSC values. In Figures 3-8c and 3-8d the SSC data are plotted as a function of wind direction (only SSC values higher than 0.2 g L\(^{-1}\) are reported). During the winter, high SSC values mostly occur during winds blowing from the east and
northeast, suggesting that sediment resuspension mostly occurs during cold front passages. During the spring, the highest SSC values occur for winds from southeast. Importantly, this is the most favorable wind direction for Mississippi River plume advection to the west along the coast (Walker et al., 2005). There are also high SSC values co-occurring with winds coming from the north-east. Winds from this direction are cross-estuary winds forcing a water level setup in response to coastal Ekman transport that facilitates estuarine-shelf exchanges.

Figure 3-8 Wave and SSC distribution during the winter and spring deployments at station BB1 (only significant wave heights higher than 0.2 m are reported). Distribution of wave height as a function of wind direction during the winter (A) and spring (B); distribution of high SSC values (higher than 0.2 g L\(^{-1}\)) as a function of wind direction during the winter (C) and spring (D).

Spectral analysis was used to explore variability in water levels, axial current velocity and SSC in the frequency domain. Spectral analysis of water levels indicated that the tidal regime in Barataria Pass was mainly diurnal (Figure 3-9a) with K1 and O1 being the dominant tidal constituents. Likewise, axial velocity in Barataria Pass exhibited predominately diurnal fluctuations (Figure 3-9b). Semidiurnal fluctuations were also present in the spectra of water level and axial velocity but were much weaker than those at diurnal frequency (note that the vertical axis in Figure 3-9 is shown on a logarithmic scale). Winter SSC values demonstrated peaks on the 23.9 and 25.8 h time scales, corresponding to K1 and O1 frequencies (Figure 3-9c), suggesting that sediment resuspension within the inlet is occurring during each tidal cycle, thus causing diurnal fluctuations in SSC data. However, the SSC spectrum from spring deployment showed the absence of a peak at diurnal time scales, likely due to the controlling influence of other factors, such as Mississippi River discharge, that imposed stronger fluctuations at lower frequencies.
Wavelet coherence analysis was used to elucidate relationships among SSC, water levels, axial velocity and wind stress in Barataria Pass (Figure 3-10). SSC values during winter were significantly coherent with water levels and axial velocity around 1-day period band. The coherency in 1-day period was always present during the spring tides when tidal range was higher but faded during the neap tides (Figures 3-10a and c). In contrast, SSC during spring was not coherent with water levels and axial velocity at 1-day period band (Figures 3-10b and d). These results are consistent with the results of spectral analysis and once again confirm that tides have the controlling influence on SSC during winter but are not a significant factor during spring. Of more interest is the connection between SSC, water level and axial velocity over longer-term periods. During the winter, a strong coherency was observed for the 3 to 10 days periods, consistently with the passage of cold fronts over the northcentral Gulf of Mexico (Figures 3-10a and c) (Payandeh et al., 2019). Although a strong coherency for the 3-10 days band also exists in spring wavelets, this coherency is less significant because of the decreased cold front activity during that time. To further analyze the connection between SSC and cold fronts, wavelet coherence was calculated among SSC and wind stress (Figures 3-10e and f). As expected, for longer periods (3 to 10 days), SSC is coherent with wind stress during both
deployments, with winter being more significant due to the higher cold front activity. Arrows in Figure 3-10e and f are more or less pointing to the right (0°), suggesting largely in phase relationship between SSC and wind stress, such that an increase in wind stress corresponds to an increase in SSC.

Figure 3-10 Wavelet coherence between sediment concentration and water level (A and B); axial velocity (C and D); wind stress (E and F) during the winter (left) and spring (right). The 5% significance level against red noise is shown as a thick contour. The horizontal axis is the number of days after the deployment. The U-shaped curves with lighter shade of the power spectrum show the cone of influence where edge effects become important. Phase arrows indicate the relative phase relationship between the series; pointing right: in-phase; left: anti-phase; down: SSC leads by 90°; up: water level/axial velocity/wind stress leads by 90°.
3.3.3 Salinity and SSC Patterns

Salinity time series were analyzed to investigate the responses of Barataria Bay to changes in the Mississippi River discharge. Salinity distribution along the bay longitudinal axis showed different patterns in winter and spring (Figure 3-11). During the winter, the salinity gradients within the bay were relatively small (Figure 3-11b). In addition, salinity values at stations BB1, BB2 and BB4 co-oscillated in phase and responded similarly to the cold front passages. During pre-frontal phase, when winds blew from south and southeast, water was pushed into the bay and salinity increased by 3-10 depending on the duration and strength of the southerly winds. Subsequently, when fronts reached the bay and winds shifted from southerly to northerly direction, all three stations experienced a sudden salinity drop of up to 15 depending on the strength of the particular front.

Contrary to the winter season, salinity gradients along the bay were significant during the spring (Figure 3-11d). Salinity values in the lower sections of the estuary (stations BB1 and BB2) co-oscillated in phase and responded similarly to the cold front winds. However, they co-oscillated out of phase with salinities of the upper bay station BB4. During prefrontal phase when winds are southerly and water is pushed into the bay, salinity at stations BB1 and BB2 dropped by 10-15. In contrast, during frontal phase when northerly winds pushed water seaward, salinity increased by 5-10 similar to an inverse estuary. Further, salinity at station BB4 increased during prefrontal phase and decreased during frontal phase. This combination of positive and negative estuary behavior indicated that estuarine salinity maximum (i.e., salt plug) was located between stations BB4 and BB2. This is consistent with the previous study of Juarez et al. (2019) who also reported salt plug formation inside Barataria Bay.
To further investigate the salt-plug formation in Barataria Bay and its connection with Mississippi River discharge, we examined the 22-year salinity data collected by Turner et al. (2019). For this study, we used a subset of data that consisted of thirteen stations inside the bay, one station in Barataria Pass and one station on the shelf (Figure 3-12). Temporal variations indicated that salinity values were the highest during the winter and the lowest during the spring. In general, the salinity inside the bay was lower than in Barataria Pass and on the shelf except when the salt plug was formed. The salt plug was observed mostly during the spring seasons and its formation was generally coincidental with the Mississippi River discharge maxima (April and May, Figure 3-12).

During the spring, increased discharge of the Mississippi River through the Bird’s Foot Delta caused mixing of freshwater with shelf waters, subsequent entrainment of less saline water
into the Louisiana Coastal Current, and its transport into Barataria Bay. This mechanism is further supported by satellite imagery (Figure 3-13). MODIS images processed using a “true color, red-green-blue” enhancement technique, confirmed the northwestward flow of the Mississippi plume water discharged from Southwest Pass and West Bay Diversion and subsequent transport into the bay. The MODIS images from March 2018 clearly show sediment-laden Mississippi River water entering the bay through the passes that carry water from the inner shelf into the bay.

In the presence of the lower salinity water in the mouth and the head of the estuary, the brackish water within the estuary transitioned into the maximum salinity zone (the salt plug). As shown in figure 3-3b, the spring deployment centered on the large discharge event (3.6x10^4 m^3 s^-1) that explains salt-plug formation during this period (Figure 3-11). Under higher SSC values due to the intrusion of the Mississippi River water into the lower Barataria Bay, tidal resuspension was no longer a controlling factor and therefore the SSC spectrum and wavelets for the spring deployment showed no coherency in diurnal frequencies (Figure 3-9c and 3-10f). The strong interaction between Mississippi discharge and sediment dynamics in Barataria Bay is further supported by a comparison of the 22-year long time series of salinity, TSS and Mississippi River discharge data (Turner et al., 2019). The comparison demonstrated a strong connection between salinity and TSS in Barataria Pass (station 2) and Mississippi River discharge (Figure 3-14). During our study period, the lowest salinity in Barataria Pass occurred during April and May (Figure 3-14a), coincidentally with the highest TSS (Figure 3-14b), and the highest Mississippi River discharge values (Figure 3-14d).
Figure 3-12 Temporal evolution of the longitudinal salinity gradients (A, B and C) as measured along the Barataria Bay by Turner et al. (2019); Mississippi River discharge (D, E and F) measured at Belle Chasse. Areas with no data available are white.
Figure 3-13 A sequence of the Terra-1 MODIS true color imagery of Mississippi River plume from March 10 to March 26, 2018. The images show a progression of a northwestward flow of Mississippi plume from SW pass and West Bay diversion and subsequent transport into Barataria Bay.
3.4 Discussion

During both deployments, resuspension by cold fronts resulted in high SSC values. Although fewer and less strong cold fronts occurred during the spring deployment, due to higher background SSC values transported from Southwest Pass of the Mississippi River during this period of time, the highest SSC value in Barataria Pass was recorded during the passage of a moderately strong cold front on March 20, 2018. Spectral and wavelet analysis showed that winter SSC had significant connection with tides. However, SSC during the spring displayed no connection with tides.

Decomposition of the total sediment flux time series revealed that subtidal advection was the primary mechanism of sediment transport during both winter and spring (Figure 3-15). Tidal pumping flux in both seasons were comparable in magnitude and much smaller than the former. During the winter the advective sediment flux was an order of magnitude larger than pumping.
flux and it was oriented down-estuary (Figure 3-15a) whereas during the spring, it was 2-5 times the pumping flux and it was oriented upstream (Figure 3-15b). This change in advective sediment flux orientation was most likely related to an increase in SSC in the BB1 area associated with westward migration of the Mississippi River plume during the second deployment. The main contribution to the advective sediment flux in the estuary is cold fronts during the winter and river flood during the spring. These data suggest that cold fronts enhanced sediment export from the bay during the winter and the advection of the river plume – compounded with the pre-frontal phases of cold fronts when winds blow from the south - enhanced sediment import during the spring. Although, our measurements were only taken at the east side of the inlet, and thus they do not reflect the whole discharge through the inlet, they are considered a proxy for the true sediment discharge through the inlet. Nevertheless, because most of the sediment is advected through the inlet (rather than being locally resuspended), even a measurement on one point is highly representative for the whole inlet. At the inlet, the peak concentration in the spring is 1.8 g L\(^{-1}\). This implies that the SSC is not simply the passive advection of the river plume. Instead, we suggest that the sediment carried by the plume during the spring is temporarily stored (i.e., for a few weeks) in the nearshore. The waves associated with cold front can eventually resuspended this sediment and lead to SSC much higher than the initial values in the river plume.

Figure 3-15 Cumulative SSC flux at station BB1 during the winter (A) and Spring (B). Positive and negative values denote landward (up-estuary) and seaward (down-estuary) transports, respectively.

The longitudinal salinity gradients during the spring season showed the existence of a salt-plug within the estuary. This is consistent with the previous study of Juarez et al. (2019) who also reported salt plug formation in Barataria Bay. Salt-plug formation has been observed
predominantly in arid climates where evaporation is typically much larger than river input. This is not the case for Barataria Bay which has a surplus of precipitation over evaporation (Das et al., 2010). Shaha and Cho (2016) described a different type of salt plug in multi-channel estuaries that is caused by decreasing river discharges. Likewise, this type of salt plug cannot develop in Barataria Bay because river discharge into Barataria Bay is generally small, which is typical of bar-built estuaries in the region. The main freshwater source for Barataria Bay is through a man-made diversion – the Davis Pond. The maximum design capacity of the Davis Pond Diversion is ~ 300 m³ s⁻¹, but the real discharge is much smaller for most of the time (Das et al., 2010). Also, the diversion is located some 85 km from Barataria Pass and so its impact on salinity in that region is negligible. In fact, Barataria Pass is much closer to the Mississippi River Southwest Pass and the West Bay Diversion on the Mississippi River than the Davis Pond Diversion (Figure 3-1). Given the high Mississippi River discharge during the spring, intrusion of the Mississippi River plume from the coastal ocean into Barataria Bay was the likely mechanism explaining high sediment concentrations during the spring. MODIS images processed using a “true color, red-green-blue” enhancement technique, confirmed the northwestward flow of the Mississippi plume water discharged from Southwest Pass and West Bay Diversion and subsequent transport into the bay. A modeling study by Zang et al., (2019) identified two transport pathways of the Mississippi-derived sediments: 1) A direct northwestward alongshore transport from the Southwest Pass; and 2) a gyre-induced clockwise transport, which joins the alongshore transport. These two pathways have been also reported by Fitzgerald et al., (2004), Georgiou et al., (2005) and Li et al., (2011). Li et al. (2009) indicated that the west side of the Barataria Pass has a much larger outward flow than the east side and the east side is fresher. They suggested that the Louisiana Coastal Current initiated on the west side of the Birdfoot Delta, comprised primarily of freshwater that may have entered Barataria Bay during the course of their study.

Based on this study, a conceptual diagram describing the SSC response of the bay to different forcing mechanisms during the winter and spring seasons is shown in Figure 3-16. During the winter, cold front and tide both control the SSC variability. The strong wind forcing increases the vertical mixing between the surface water and the subsurface water, and the tidal mixing significantly enhances the sediment resuspension in the subsurface and bottom water and thus increases the SSC values during the winter. During the spring, SSC is primarily regulated by the freshwater discharge of the Mississippi River and secondarily by the mixing in response to cold front passages. High Mississippi discharge during spring causes mixing of the river plume with the Louisiana Coastal Current. Subsequently, turbid waters enter the bay through several tidal inlets, which combined with strong cold front winds further enhances vertical mixing and increases SSC.
3.5 Conclusion

This study provides new insights into the SSC dynamics of complex bar-built estuaries affected by offshore river plume dynamics. Two series of opportunistic observations captured the coupled estuarine-shelf interactions within the main inlet of Barataria Bay. During the winter, the SSC dynamics was controlled by resuspension driven by both cold fronts and tides. During the spring, as the Mississippi River plume progressed along the Louisiana coast, shelf influences began to strongly affect the SSC responses within the estuary via a surface advected river plume. The results of this study further confirmed previous findings suggesting that the Mississippi River water can at times enter the bay through Barataria Pass. Importantly, this study also showed the strong connection between the intrusion of the Mississippi River water and salt plug formation in Barataria Bay. As a result of salt plug formation, inverse estuarine circulation prevails upstream and downstream of the salinity maximum zone and therefore the salt plug acts as a barrier inhibiting the mixing of shelf and estuarine waters. Such an inverse circulation has important consequences for sediment transport because the sediments discharged into the estuary from upstream sources will remain within the estuary until the salt plug is dissipated. Collectively, these findings advance our ability to predict how bar-built estuaries respond to temporal changes of river discharge and offshore river plume dynamics. Overall, we conclude that the sediment discharged at the mouth of the Mississippi River is not completely wasted, but instead partly supplies Barataria Bay through its seaward boundary. Future investigations should better quantify this contribution, for example by measuring fluxes through the whole inlet and through the other inlets of Barataria Bay and compare this offshore sediment input to the direct sediment input by river diversions directly into the bay.
CHAPTER 4. TIDAL CHANGE IN RESPONSE TO THE RELATIVE SEA LEVEL RISE AND MARSH ACCRETION IN A TIDALLY CHOKE ESTUARY

4.1 Introduction

Tidal range is arguably the most important parameter determining the hydrodynamics of estuaries and coastal embayments. Accelerated sea level rise has the potential to affect tidal range because sea level rise changes coastlines, increases water depth, and floods low-lying land in estuaries, thereby changing the balance between the bottom friction and other forces. Changes in tides may have important impacts on ecosystem dynamics of estuaries. For instance, tidal prism will be affected by changes in tidal range and, hence, the flushing capacity and residence time of an estuary will be affected as well (Du & Shen, 2016; Monsen et al., 2002). Additionally, tidal changes are often associated with shifts in sediment transport, salinity intrusion, and ecosystem properties (Talke & Jay, 2020).

Tidal range response to sea level rise is complex, and has been showed to either increase or decrease, and change along the estuary axis. Multiple modeling studies have examined this response. Hagen and Bacopoulos (2012) compared the static versus dynamic response of tides to sea level rise and reported that changes in astronomic tides should be assessed as a dynamic process and not as a static process. Passeri et al., (2015a) evaluated the combined effects of historic sea level rise and morphology changes on tidal hydrodynamics in Grand Bay, Mississippi. They demonstrated that tidal amplitudes significantly increased in the semi-enclosed regions, however changes within the sound were minimal due to the open exposed shoreline. Lee et al., (2017) developed a numerical model to investigate how sea level rise may impact tides in Chesapeake Bay and Delaware Bay. They reported similar responses in both estuaries and found that when the low-lying land is prevented from flooding, tidal range increases and when it is allowed to become permanently inundated by higher sea level, tidal range decreases. Likewise, Holleman & Stacey (2014) implemented a numerical model to study tidal changes in San Francisco Bay under future sea level rise scenarios. They found that inundation of low-lying areas introduces frictional regions that serves as energy sinks for tides, thus decreasing the tidal range under higher sea levels. Despite these similarities, tidal response to sea level rise varies among different estuaries depending on the estuarine characteristics. Du et al., (2018) found that tidal range changes under sea level rise are heavily dependent on an estuary’s length and bathymetry. They showed that estuaries with a narrow channel and large low-lying areas are likely to experience decreased tidal ranges under higher sea levels. Tides increase differently in various bays in the Northern Gulf of Mexico for different sea level rise scenarios in ways related to the projected change in inlet area (Passeri et al., 2016). Talke & Jay (2020) identified two types of systems that are prone to tidal amplification: (1) shallow, strongly damped systems, in which a small increase in water depth induces a large decrease in friction, and (2) systems in which wave reflection and resonance are strongly influenced by changes in depth, convergence and friction.

Tidal response to sea level rise is also sensitive to morphological changes (Bilskie et al., 2014; Passeri et al., 2015a). However, due to uncertainties in predicting the future geomorphology, standard approaches to exploring the impacts of sea level rise on tidal dynamics do not consider geomorphology changes (e.g., Holleman & Stacey 2014; Lee et al., 2017). Passeri et al., (2016) used a numerical model coupled with a probabilistic model that projects
shoreline change and dune heights under future sea level rise conditions to examine the combined effects of sea level rise and morphology changes on tidal hydrodynamics along the Northern Gulf of Mexico. Changes in tidal range may also affect the stability of coastal marshland as it determines the ability to transfer sediment to the marsh platform and thus to accrete (Kirwan & Brad Murray 2007). In addition to biological and physical sediment transport processes, the ability of marshes to maintain their position depends on tidal range (Friedrichs and Perry, 2001; Torres et al., 2006). Tidal range indicates the frequency and duration of tidal inundation and thus affects rates of mineral sediment deposition on marshes (Pasternack et al., 2000). Generally coastal marshes have kept their morphology under historic rates of sea level rise (Redfield, 1972). However, in recent decades, their ability to survive with more rapid rates of sea level rise has been questioned (Orson et al., 1985; Reed, 1995). Mariotti and Fagherazzi, (2010) showed that if the rate of sea level rise is low, wave dissipation is enhanced which favors sediment deposition, thus the marsh boundary progrades. In contrast, a high rate of sea level rise leads to higher waves, leading to erosion of marsh. When the rate of sea level rise is too fast the entire marsh drowns and is transformed into a tidal flat. Hagen et al., (2013) studied the impact of sea level rise on salt marsh sustainability in the lower St. Johns River. They found that mean low water and mean high water are both sensitive to sea-level rise and increase by an amount that is not proportional to the level of sea level rise. Hence in numerical modeling studies, it is important to update the morphology based on the future sea level rise conditions.

Coastal Louisiana is experiencing some of the highest sea level rise rates in the US, and thus is ideal for examining the impacts of sea level rise on estuarine tidal dynamics. In recent decades, this region has experienced globally high rates of land loss resulted from a number of anthropogenic and natural factors. Land loss in Louisiana is the result of a variety of factors, including sea level rise, subsidence, man-made disturbances, wave erosion, and saltwater intrusion, among others, but many researchers point to the relative sea level rise (RSLR = Eustatic Sea Level Rise (ESLR) + subsidence) as a major driver of land loss (Day et al., 2000; Day et al., 2007; Paola et al., 2011; Hiatt et al., 2019). The total land area lost in Mississippi River delta over the last 100 years has been approximately 5000 km² at rates as high as 100 km² yr⁻¹ (Day et al., 2000; Couvillion et al., 2017). Here we focus on Barataria Basin (also referred to Barataria Estuary), which is located between the main stem of the Mississippi River and Bayou Lafourche (Figure 4-1). Barataria Basin has unique features that distinguish it from other estuaries studied before. Barataria Basin is a very shallow estuary (averaged depth ~ 2 m) which features a complex geometry involving many lakes, bayous, channels, tidally influenced marshes and barrier islands on the south that separate the whole system from the Gulf of Mexico. There are large regions of freshwater, brackish, and saline marshes that account for more than 60% of total area of the bay. These low-lying wetlands and marshes are prone to flooding when sea level rises. Tides in Barataria Basin are microtidal, with K₁ and O₁ being the dominant tidal constituents, whose amplitudes in Barataria Pass are about 15 cm. Barataria Basin is a tidally chocked system and dissipates tides heavily so that they are attenuated by 68% at Lafitte located at the middle of the estuary (Figure 4-1; Byrne et al., 1976; Conner et al., 1987). The combined effects of RSLR and potentially increasing tidal ranges will have pervasive effects on Barataria Basin. Thus, as sea level rises, it is imperative to quantify how tidal dynamics will be affected by RSLR.

Here we investigate the following questions: (i) How tidal range will be affected by RSLR (ii) How amplitudes and phases of individual tidal harmonics will be affected (iii) How tidal inundation extents will be affected by RSLR and changing tides (IV) How the forcing
balance in the estuary will change under higher sea levels and (V) how the vertical accretion of marsh changes the response to all the previous questions.

Figure 4-1 The map of the (a) northern Gulf of Mexico and (b) Barataria Basin showing bathymetry and location of the stations used in momentum balance analysis (M1, M2, and M3)
4.2 Methodology

A numerical model was developed to explore the effects of RSLR and marsh accretion on tidal dynamics in Barataria Basin. Different scenarios were conducted to simulate tides under various future RSLR and marsh accretion conditions. The changes in the tidal dynamics were explored through an analysis of the momentum equation.

4.2.1 Model Setup

In this study, the Finite Volume Community Ocean Model (FVCOM) was used to explore the effects of RSLR on tidal dynamics. FVCOM solves the governing equations using an unstructured grid (Chen et al., 2003, 2013). It has been used successfully in many studies around the world including studies of the northern Gulf of Mexico (Huang et al., 2011; Huang & Li, 2017; Li et al., 2011; Payandeh et al., 2019; Wang & Justic, 2009). The system of governing equations consists of the momentum, continuity, temperature, salinity, and density equations under the Boussinesq and hydrostatic approximation. The detailed formulations and numerical aspects of FVCOM can be found in Chen et al., (2013) and they will not be repeated here.

Sea level rise could change tidal amplitudes in the open ocean and over the continental shelf (Pickering et al., 2012, 2017). Therefore, FVCOM was implemented to the entire Gulf of Mexico with a focus on the northern gulf coast and particularly a fine spatial resolution (~15 m) in Barataria Basin. As shown in Figure 4-2, the model domain covers all the lakes, bayous, channels and major waterways along the estuary. The numerical mesh consists of 700,224 cells and 356,816 nodes. Bathymetric data were obtained from Coastal Louisiana Ecosystem Assessment and Restoration Report (CLEAR), and Digital Elevation Models (DEM) developed by NOAA's National Geophysical Data Center (NGDC) with 10 m resolution and were interpolated to the grid using an inverse distance weighted method. The model has two open boundaries as shown in Figure 4-2: (i) a latitudinal line in the northern Caribbean Sea and (ii) the Strait of Florida. The tidal forcing was implemented by specifying the amplitudes and phases of eight dominant astronomic tides ($S_2$, $M_2$, $N_2$, $K_2$, $K_1$, $O_1$, $P_1$, and $Q_1$) at model open boundaries, derived from ADCIRC EC2015 tidal databases (Szpiłka et al., 2016). Tidal potential forcing was also applied in the Gulf of Mexico using the same eight tidal constituents. The flooding and drying treatment was implemented to take into account the water level fluctuations over intertidal zone and allow the flooding of low-lying lands beyond the present shorelines as sea level rises. All simulations began with a null state and were run for 100 days. The first 20 days were model warm-up period, over which all forcings were ramped up from zero to their full values. Thus, all harmonics were analyzed during the last 80 days. Future sea level rise scenarios were implemented in the model by adding an additional steady component to open boundaries. This component had a zero-phase and an amplitude equal to the ESLR for the given scenario. In addition, in each scenario, bathymetry was updated to represent the subsidence and marsh accretion in the future conditions. Changes in tidal range were quantified by calculating the Mean Tidal Range (MTR) as well as the amplitude of individual harmonics such as $K_1$ and $O_1$. MTR was calculated as: mean high water (MHW) - mean low water (MLW). MTR difference was also calculated as the MTR in each scenario minus MTR under the present sea level condition.
4.2.2 Model Experiments

Siverd et al. (2019) demonstrated the need to take into account ESLR and subsidence (i.e., RSLR) when examining future hydrodynamic responses on the Louisiana coastal land margins. A numerical modeling study by Blum et al., (2008) suggested subsidence rates of approximately 1 mm yr$^{-1}$ produced by lithospheric flexure. Veatch et al., (2015) reported a relatively long time series (~ 50 years) of RSLR values derived from 19 tide gauges in Louisiana Coast. They reported a RSLR rate of 10.8, 7.0, and 11.75 mm yr$^{-1}$ in Bayou Lafourche, Bayou Barataria, and West Pointe a la Hache respectively. Hence, subsidence rates are spatially different and such complexity made it challenging to define values for the future RSLR scenarios. In this study nine distinct RSLR scenarios categorized into three major classes of Lowest, Medium and Highest were considered for the next 50 years. Subsidence and ESLR in each scenario were defined following a similar approach developed by the 2012 Coastal Master Plan (CPRA, 2012). Table 4-1 provides an overview of the RSLR scenarios implemented in this study that include, ESLR values, subsidence ranges and the marsh accretion rates considered in each scenario. Subsidence ranges were derived from a map of plausible subsidence rates (ranging
from 0 to 35 mm yr\(^{-1}\)) for coastal Louisiana, separated into 17 geographical regions (CPRA, 2012- Appendix C). For example, the lower Barataria Basin subsidence zone had a range of observed subsidence rates between 6 and 20 mm yr\(^{-1}\), therefore, in the highest scenario, the 50th percentile value is equal to 13 mm yr\(^{-1}\) or 13*50=650 mm/50 years.

Incorporating coastal mechanisms that adjust morphology for defining future inundation under RSLR and informing the model is challenging due to the lack of consistent models that predict accretion and erosion in response to sea level rise (Zhang, 2011; Passeri et al, 2015b). Short term accretion rates provided by Coastwide Reference Monitoring System (CRMS) show high spatial variability in Barataria Basin from 0.48 cm yr\(^{-1}\) in Bayou Wilkinson to 2.39 cm yr\(^{-1}\) in Bayou Dupont. Comparison of long-term accretion rates produced by Shrull, (2018) and CRMS short term data show that these rates are time dependent, i.e., the period of observation can change the observed rates. Here, in order to explore the effect of marsh accretion on tidal dynamics and as a first order approximation, three different subsidence rates are considered for each RSLR scenario: (i) without accretion, in which the marsh has no vertical accretion in response to the RSLR, (ii) an accretion rate equal to 50% of its corresponding RSLR value, and, (iii) an accretion rate equal to 100% of its corresponding RSLR value, where the marsh is able to keep pace with RSLR (Table 4-1). Louisiana has followed a comprehensive coastal protection and restoration plan that seeks to protect all of the barrier islands in the hopes of minimizing damage from future sea level rise and hurricanes. So, in this study it is assumed that the present shorelines around barrier islands of the Barataria Basin will not change, thus water is not allowed to flood the barrier islands but flooding of extensive wetlands inside and around the estuary is allowed.

Table 4-1 Values used in the nine future RSLR Scenarios for the next 50 years (RSLR=ESLR +Subsidence). Subsidence ranges were derived from a map of plausible subsidence rates for coastal Louisiana, separated into 17 geographical regions (CPRA, 2012- Appendix C).

<table>
<thead>
<tr>
<th>Scenario</th>
<th>ESLR (m/50 yr)</th>
<th>Subsidence (m/50 yr)</th>
<th>Accretion rate of marsh (m/50 yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lowest1</td>
<td>0.43</td>
<td>20% of range</td>
<td>without accretion</td>
</tr>
<tr>
<td>Lowest2</td>
<td>0.43</td>
<td>20% of range</td>
<td>50% of RSLR</td>
</tr>
<tr>
<td>Lowest3</td>
<td>0.43</td>
<td>20% of range</td>
<td>100% of RSLR</td>
</tr>
<tr>
<td>Medium1</td>
<td>0.63</td>
<td>20% of range</td>
<td>without accretion</td>
</tr>
<tr>
<td>Medium2</td>
<td>0.63</td>
<td>20% of range</td>
<td>50% of RSLR</td>
</tr>
<tr>
<td>Medium3</td>
<td>0.63</td>
<td>20% of range</td>
<td>100% of RSLR</td>
</tr>
<tr>
<td>Highest1</td>
<td>0.83</td>
<td>50% of range</td>
<td>without accretion</td>
</tr>
<tr>
<td>Highest2</td>
<td>0.83</td>
<td>50% of range</td>
<td>50% of RSLR</td>
</tr>
<tr>
<td>Highest3</td>
<td>0.83</td>
<td>50% of range</td>
<td>100% of RSLR</td>
</tr>
</tbody>
</table>

4.2.3 Error Analysis

At 30 stations, the simulated 80 days water levels were compared with the reconstructed tidal water levels, i.e., the water levels constructed by using the eight dominant harmonic constituents for each station. Amplitudes and phases of the harmonic constituents were either derived from NOAA website (https://tidesandcurrents.noaa.gov/) or calculated from observed USGS water levels. In the rest of the analysis, we will refer to the reconstructed water levels as observed water levels. In addition, tidal constituent amplitudes and phases were computed and
compared at each station using two metrics. The first metric is the standard deviation (STD), which is calculated as the root mean square error:

\[
STD = \sqrt{\frac{1}{N-1} \sum_{i=1}^{N} (OBS_i - SIM_i)^2}
\]

(Eq. 4.1)

where OBS is the observed value and SIM is the simulated value. The second metric is the correlation coefficient squared \((R^2)\) which is calculated as the square of correlation coefficient (Thomson and Emery, 2014):

\[
R = \frac{1}{N-1} \sum_{i=1}^{N} \frac{(OBS_i - \overline{OBS})(SIM_i - \overline{SIM})}{STD_{OBS} STD_{SIM}}
\]

(Eq. 4.2)

where OBS is the mean of the observed values, SIM is the mean of the simulated values, STD_{OBS} is the standard deviation of observed values and STD_{SIM} is the standard deviation of simulated values.

4.3 Results

4.3.1 Model-observation comparisons

The simulated tidal water levels for the present scenario were compared with observed tidal water levels obtained from NOAA tide gauges in 30 stations. These stations are listed in Table 4-2 and span from Key West in Florida to Brazos Island in Texas. Of these stations, 10 are located in Barataria Basin and span from Grand Isle in the south to Bayou Gauche in the north. An example of model-observation comparisons is given in Figure 4-3 which shows 60-days long comparisons at 10 stations inside Barataria Basin. The simulated water levels are directly extracted from the FVCOM output and the observed water levels are those constructed from the eight primary tidal constituents \((S_2, M_2, N_2, K_2, K_1, P_1, O_1, Q_1)\). As a quantitative measure of model accuracy, the correlation coefficients squared, \(R^2\) and standard deviation, STD between simulated and observed water levels are computed. Overall, the simulated water levels were in good agreement with the observed water levels. \(R^2\) ranged from 0.91 to 0.98 and the STD ranged from 0.005 to 0.05 m. Higher resolution in Barataria Basin resulted in higher accuracy in the study area where STD ranged from 0.005 to 0.02 m.

To further examine the efficiency of the model, a comparison is made between the NOAA/USGS-measured and the FVCOM-computed amplitudes and phases of tidal constituents at 30 stations. A comparison for the 4 dominant constituents \((M_2, K_1, P_1, O_1)\) in Barataria Basin and the whole domain, respectively is presented in Figure 4-4. Different bands are defined at 0.025 and 0.05 m for the amplitude plots and 10° and 20° for the phases plot. Most of the amplitudes fall inside the 0.025 error band and others fall very near or inside the 0.05 m error band. For phases, most of the constituents fall in 10° error band and others in 20° error band. Only some phases of \(M_2\) show significant differences. Semidiurnal constituents like \(M_2\) are more dominant in Florida shelf up to Apalachicola. Additional resolution in this region may be required to improve the semidiurnal prediction. In addition, the NOAA measured data include measurement uncertainties due to changing bathymetry of coastal regions and nontidal events including river discharges, wind-driven events, and radiational heating cycles. These
uncertainties can account for 35%–60% of the modeled to observed amplitudes and 50–80% of the phase difference (Bunya et al., 2010). Nevertheless, the results were satisfactory. The squared correlation ($R^2$) was 0.94 and 0.96 for amplitudes and phases in the whole domain. In Barataria Basin the squared correlation was 0.90 and 0.96 for amplitudes and phases, respectively. The standard deviation of amplitudes was less than 0.02 m and the standard deviation of phases was less than 20° everywhere. Overall, these results indicate an excellent match between simulated and observed amplitudes and phases.

Table 4-2 List of 30 stations used in the model-observation comparisons.

<table>
<thead>
<tr>
<th>State</th>
<th>Station ID</th>
<th>NOAA/USGS ID</th>
<th>Station location</th>
</tr>
</thead>
<tbody>
<tr>
<td>TX</td>
<td>SLR01</td>
<td>8775270</td>
<td>Port Aransas</td>
</tr>
<tr>
<td>TX</td>
<td>SLR02</td>
<td>8772471</td>
<td>Freeport Harbor</td>
</tr>
<tr>
<td>LA</td>
<td>SLR03</td>
<td>8768094</td>
<td>Calcasieu Pass</td>
</tr>
<tr>
<td>LA</td>
<td>SLR04</td>
<td>8763535</td>
<td>Caillou Bay</td>
</tr>
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<td>LA</td>
<td>SLR05</td>
<td>8762888</td>
<td>Lake Pelto</td>
</tr>
<tr>
<td>LA</td>
<td>SLR06</td>
<td>8762223</td>
<td>Timbalier Bay</td>
</tr>
<tr>
<td>LA</td>
<td>SLR07</td>
<td>8761724</td>
<td>Grand Isle</td>
</tr>
<tr>
<td>LA</td>
<td>SLR08</td>
<td>291929089562600</td>
<td>Grand Terre Island</td>
</tr>
<tr>
<td>LA</td>
<td>SLR09</td>
<td>8761742</td>
<td>Mendicant Island</td>
</tr>
<tr>
<td>LA</td>
<td>SLR10</td>
<td>8761819</td>
<td>Hackberry Bay</td>
</tr>
<tr>
<td>LA</td>
<td>SLR11</td>
<td>292859090020040000</td>
<td>S of Lafitte</td>
</tr>
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<td>LA</td>
<td>SLR12</td>
<td>2928000900600000</td>
<td>Bay Dosgris</td>
</tr>
<tr>
<td>LA</td>
<td>SLR13</td>
<td>07380335</td>
<td>Little Lake</td>
</tr>
<tr>
<td>LA</td>
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<td>Bayou Perot</td>
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<tr>
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</tr>
<tr>
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</tr>
<tr>
<td>LA</td>
<td>SLR17</td>
<td>8760943</td>
<td>SW Pass</td>
</tr>
<tr>
<td>LA</td>
<td>SLR18</td>
<td>8760551</td>
<td>South Pass</td>
</tr>
<tr>
<td>LA</td>
<td>SLR19</td>
<td>8760417</td>
<td>Devon Energy Facility</td>
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<td>LA</td>
<td>SLR20</td>
<td>8760668</td>
<td>Grand Pass</td>
</tr>
<tr>
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<td>SLR21</td>
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<td>SLR29</td>
<td>8724580</td>
<td>Key West</td>
</tr>
<tr>
<td>FL</td>
<td>SLR30</td>
<td>8724698</td>
<td>Loggerhead Key</td>
</tr>
</tbody>
</table>
Figure 4.3 Comparisons between modeled and observed tidal water levels at 10 sample stations. Station names are shown at the top right of the panels. The squared correlation coefficient ($R^2$) and standard deviations ($STD$) in meters are shown in the plots.
Figure 4-4 Comparisons between simulated and observed tidal constituents (amplitudes and phases) for the 4 dominant constituents (M$_2$, K$_1$, P$_1$, O$_1$) in 30 stations. Results are separated by Barataria Basin and the whole domain. The squared correlation coefficient ($R^2$) and standard deviations ($STD$) in meters are shown in the plots. The error bands are defined at 0.025 and 0.05 m for the amplitude plots and 10° and 20° for the phases.
Figure 4-5 Tidally induced inundation of the estuary under the present sea level condition and future RSLR scenarios. Inundation is shown in terms of the percent of the simulation time that a specific area is flooded, e.g., 100% means permanently flooded and 0% means never flooded.
Figure 4-6 Mean tidal range under the present sea level condition and future RSLR scenarios.

4.3.2 Tidal dynamics under present sea level conditions

Under the present sea level conditions, the extent of the flooded areas in each scenario is shown in Figure 4-5 and summarized in Table 4-3. At present, 1787 km$^2$ of the estuary was permanently inundated which is 42% of the total estuary area and the estuary had a very complex geometry with a vast intertidal area. Although the lower estuary was relatively open, the upper estuary comprises mostly divided water bodies that were only connected by small bayous and channels. Barataria Waterway which is a relatively deep shipping channel (~4 m) and runs from the Barataria Pass along the axis of the estuary and extends to the Gulf Intracoastal Waterway, had an important role in connecting divided water bodies (Figure 4-1).

$K_1$ and $O_1$ were the dominant tidal constituents, whose amplitudes on the shelf were about 15 cm (Figures 4-7a, and 4-8a). These amplitudes were constant on the shelf but sharply decreased when passing through Barataria Pass where they dropped to 9 cm just inside the bay. Five major tidal choking areas were identified within the bay represented by the vertical gray
dashed lines in Figure 4-7. The common feature in all of these areas was a narrow pass that connects two relatively larger water bodies within the estuary. The first tidal choking occurred in Barataria Pass as described above. In Barataria Pass, in addition to amplitude dissipation, up to 2.5 hr of phase difference was also developed (Figures 4-7b and 4-9a). In the lower estuary, amplitudes and phases of $K_1$ remained relatively unchanged around 10 cm and 4 hr respectively. The second tidal choking occurred when passing from the lower estuary into the Little Lake. Four small bayous connect the lower estuary to Little Lake including Bayou Saint Denis, Bayou Dosgris, Grand Bayou and Snail Bayou. The amplitude of $K_1$ was reduced by 3 cm in this area and the phase was lagged by 2.25 hr. The next tidal choking area was near cutoff where a small inlet connects Little Lake to Bayou Perot. 3 cm of amplitude decay and 1.2 hr of phase lag was developed at this point. Tidal dissipation continued along the Bayou Perot since it is a very shallow (~1m) and relatively narrow bayou. The next major tidal choking occurred in a narrow pass connecting Bayou Perot to Lake Salvador where amplitudes were decreased by 1 cm and phases were lagged by 2.8 hr. Lake Salvador is relatively deep (~3-5 m) and thus tidal amplitudes and phases remained relatively unchanged in this body of water. The last major tidal choking occurred when tides in Lake Salvador propagated into Bayou des Allemands. In this narrow bayou, amplitudes were reduced by 1.5 cm and phases were lagged by 8.2 hr. In Lac des Allemands located at the end of the domain, $K_1$ amplitude was almost zero and its phase was about 21 hr. Tidal phase difference between the mouth and the head of the estuary for the $K_1$ harmonic was 19.78 hr (Table 4-3).

MTR was largest (33 cm) on the shelf and diminished drastically right after the Barataria Pass (22 cm) (Figures 4-6a and 4-7g). MTR was relatively constant around 22 cm in the lower bay and started diminishing again when passing from the lower estuary into the Little Lake. As expected, the MTR changes along the estuary showed the same pattern as $K_1$ amplitudes explained above. After five major tidal chokings, the MTR reached a minimum value of 2 cm in Lac des Allemands. Thus, the MTR difference between the mouth and the head of the estuary at the present sea level was 31 cm (Table 4-3).

### 4.3.3 Response to RSLR

RSLR brought negligible tidal changes on the shelf side of the bay. However, it made substantial changes to the tidal range and tidal harmonics within the estuary (Figure 4-6). An overview of MTR and phase difference between the mouth and the head of the estuary as well as inundation extent in each scenario is given in Table 4-3. In each group of simulations (Lowest, Medium, and Highest) the most extensive inundation occurred, as expected, when no accretion was considered i.e., in the Lowest1, Medium1 and the Highest1 scenarios (Figures 4-5b, 4-5e and 4-5h). The largest inundation among all scenarios occurred in the Highest1 which was the less optimistic scenario with the largest ESLR and subsidence and no accretion of marshes. In this scenario, 3206 km$^2$ which accounts for 75 % of the total estuary area was flooded permanently i.e., was flooded more than 99% of simulation time. The remaining 25 % of the estuary area was also partially flooded (10 to 99 % of the simulation time). Even in the Lowest2, Medium2 and Highest2 scenarios in which marsh area was accreted by 50% of RSLR, extensive low-lying lands were flooded permanently (Figures 4-5c, 4-5f and 4-5i). For example, in the Highest2 case 2014 km$^2$ was permanently flooded, equivalent of 48 % of total estuary area and 2149 km$^2$ was partially flooded equivalent of 51% of total estuary area. As expected, the inundation extent in the Lowest3, Medium3 and highest3 scenarios were similar to the present
condition as it was assumed that marsh keeps pace with RSLR and therefore it was accreted by 100% of RSLR (Figures 4-5d, 4-5g and 4-5j).

Barataria Basin experienced spatially uneven change of tidal range under RSLR and the most significant changes in each scenario occurred in newly flooded areas as they were dry in the present case but flooded by RSLR (Figure 4-6). Beyond the newly flooded areas, tidal changes were smaller on the estuary’s main stem that was already flooded in the present case. In these areas and in simulations with no accretion (Lowest1, Medium1 and Highest1) changes in MTR were higher at the head of the estuary compared to the lower and middle estuary (Figures 4-6b, 4-6e, 4-6h). MTR under the Lowest1, Medium1 and Highest1 scenarios at the lower and middle estuary increased by a maximum of 2.0, 3.0 and 4.0 cm respectively. However, the maximum increase in MTR at the head of the estuary for the same scenarios were 3.0, 6.0 and 13 cm (Figure 4-7h). In contrast, in simulations with 100% of RSLR accretion (Lowest3, Medium3, and Highest3) tidal changes in the lower and middle estuary were higher than the upper estuary (Figures 4-6d, 4-6g, 4-6j). MTR under the Lowest3, Medium3 and the Highest3 scenarios at the lower and middle estuary increased by a maximum of 10, 12 and 16 cm, respectively. However, MTR at the head of the estuary for the same scenarios were increased by a maximum of 2.0, 2.5 and 4 cm respectively (Figure 4-7h).

Table 4-3 K1 phase difference ($\Delta \varphi$($K_1$)) and MTR difference ($\Delta \eta$) between the mouth (Barataria Pass) and the head (Lac des Allemands) of the estuary and tidally induced inundation extents at the present and each RSLR scenario.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>$\Delta \varphi$($K_1$)(hr)</th>
<th>$\Delta \eta$(m)</th>
<th>Area Undated (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>more than 99 % of the simulation time</td>
</tr>
<tr>
<td>Present</td>
<td>19.78</td>
<td>0.31</td>
<td>1787</td>
</tr>
<tr>
<td>Lowest1</td>
<td>17.28</td>
<td>0.29</td>
<td>2679</td>
</tr>
<tr>
<td>Lowest2</td>
<td>17.29</td>
<td>0.30</td>
<td>2023</td>
</tr>
<tr>
<td>Lowest3</td>
<td>16.61</td>
<td>0.29</td>
<td>1999</td>
</tr>
<tr>
<td>Medium1</td>
<td>15.12</td>
<td>0.25</td>
<td>2662</td>
</tr>
<tr>
<td>Medium2</td>
<td>17.17</td>
<td>0.30</td>
<td>2000</td>
</tr>
<tr>
<td>Medium3</td>
<td>16</td>
<td>0.28</td>
<td>1982</td>
</tr>
<tr>
<td>Highest1</td>
<td>11.89</td>
<td>0.17</td>
<td>3206</td>
</tr>
<tr>
<td>Highest2</td>
<td>17.11</td>
<td>0.29</td>
<td>2014</td>
</tr>
<tr>
<td>Highest3</td>
<td>14.95</td>
<td>0.27</td>
<td>1982</td>
</tr>
</tbody>
</table>

Tidal dissipation was the highest under the Lowest2, Medium2 and Highest2 scenarios because accretion of marsh by 50% of RSLR introduced extensive intertidal areas to the estuary that served as a sink for tidal energy. On the other hand, tidal dissipation was the lowest under the Lowest1, Medium1 and Highest1 scenarios because the flooding of low-lying areas turned the estuary into a widely open water body and significantly reduced tidal choking and frictional effects. MTR difference between the mouth and the head of the estuary clearly shows this effect. For example, the MTR difference between the mouth and the head of the estuary in the Highest2

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and Highest3 were 29 and 17 cm respectively. In addition, tidal phase difference between the mouth and the head of the estuary for K1 harmonic also reconfirmed the above-mentioned pattern. For example, phase difference between the mouth and the head of the estuary under the Highest2 and Highest3 scenarios were 17.11 hr and 11.89 hr respectively (Table 4-3).

Simulation results demonstrated amplification of tides at the head of the estuary. This amplification was absent under the present condition and highest under the Highest1 scenario (Figure 4-7g). Under the Highest1 scenario, tidal range reached a minimum at the middle of the Bayou des Allemends approximately 92 km from the estuary’s mouth and was amplified substantially (5 cm) at the head of estuary. This amplification can also be seen in K1 amplitudes (Figures 4-7 and 4-8). MTR difference was never negative at any of the scenarios showing that the RSLR only increased the tidal range in Barataria Basin (Figure 4-6).

Figure 4-7 Variation in K1 amplitude, MTR and MTR difference along the estuary starting 8 km off the Barataria Pass. Each vertical gray dashed line represents a narrow pass that connects important water bodies within the estuary.

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Figure 4-8 K1 amplitudes under the present sea level condition and future RSLR scenarios.
4.3.4 Analysis of the momentum equation

To further investigate how forcing mechanisms change under the higher sea levels, a momentum balance analysis was done using the vertically averaged momentum equations with constant density (Chen et al., 2013):

$$\frac{1}{D} \frac{\partial U}{\partial t} = -\frac{1}{D} \left( \frac{\partial U \partial D}{\partial x} + \frac{\partial V \partial D}{\partial y} \right) + f V - g \frac{\partial \zeta}{\partial x} \frac{\tau_{bx}}{D \rho_0} + F_x + \frac{1}{D} G_x$$

(Eq. 4.3)
\begin{equation}
\frac{1}{D} \frac{\partial \bar{V}}{\partial t} \left. = \frac{1}{D} \left( \frac{\partial \bar{U}}{\partial x} + \frac{\partial \bar{V}}{\partial y} \right) \right. - \bar{\rho} \frac{\partial \bar{z}}{\partial y} \frac{\partial \bar{y}}{\partial \rho_0} \frac{\bar{F}_y}{\bar{V}} + \frac{1}{D} \bar{G}_y + \frac{1}{D} \bar{G}_y \right)
\end{equation}

(Eq. 4.4)

where all variables are conventional, and the overbars denote the vertical integration. The terms from left to right in Equations 4.3 and 4.4 are local acceleration (DDT), nonlinear advection (ADV), Coriolis force (COR), barotropic pressure gradient (DP), bottom friction (FRIC), 2-D horizontal viscosity (VIS), and the difference between nonlinear terms of vertically averaged 2-D variables and vertical integration of 3-D variables (AV2D). The expressions for \( \bar{F}_x \), \( \bar{F}_y \), \( \bar{G}_x \) and \( \bar{G}_y \) can be found in Chen et al., (2013).

Time series of the terms in Equations 4.3 and 4.4 for three stations located at the mouth (M1), in the middle (M2) and upper estuary (M3) and for three scenarios of Highest1, Highest2 and Highest3 are shown in Figures 4-10 and 4-11 (see Figure 4-1 for exact stations location).

Figure 4-10 Time series of vertically averaged momentum equation terms in x direction at three stations for the present sea level condition and Highest1, Highest2, and Highest3 RSLR scenarios. DDT represents the local acceleration, FRIC the bottom friction, DP the barotropic pressure gradient, COR the Coriolis force, VIS the horizontal viscosity, and ADV the nonlinear advection. \( U \) is x velocity component, which illustrates the tidal cycle. See Figure 4-1 for the exact location of the M1, M2 and M3 stations.
Under the present sea level condition and in Barataria Pass (station M1), the maximum value of positive $u$ (1.07 m s$^{-1}$) was greater than the maximum negative $u$ (0.78 m s$^{-1}$). Also, the maximum value of positive $v$ (0.81 m s$^{-1}$) was greater than the maximum negative value of $v$ (0.71 m s$^{-1}$) (Figures 4-10a and 4-11a). It indicates that tidal velocity was asymmetric in Barataria Pass. Under the Highest1 simulation, both positive and negative $u$ velocities increased in magnitude, but the $v$ velocity remained unchanged. The maximum positive $u$ increased to 1.18 m s$^{-1}$ and maximum negative $u$ increased to 0.93 m s$^{-1}$ (Figures 4-10b and 4-11b). These velocities remained unchanged under the Highest2 scenario (Figures 4-10c and 4-11c) but under the Highest3 simulation, the $u$ and $v$ velocities were very close to their present values (Figures 4-10d and 4-11d). Thus, the asymmetric nature of the currents was remained with RSLR, but its intensity weakened in the Highest1 scenario. The main momentum balance at station M1 was between DP and ADV. This was true for both flood and ebb cycle and for all RSLR scenarios. This is consistent with the previous study of Cui et al., (2018) who also reported the same balance between DP and ADV in Barataria Pass using a three-dimensional baroclinic FVCOM model. The above indicates that the tidal phenomenon at this location (depth being ~20 m) is dominated by wave dynamics, as pointed out by Huang et al., (2011) and Cui et al., (2018). Both DP and ADV grow in magnitude as sea level rises and the balance between them is remained unchanged.
In the mid-estuary and under the present sea level, the \( u \) velocity was greater than the \( v \) velocity because the M2 station is located in a bayou oriented mainly in an east-west direction (Figures 4-10e and 4-11e). The depth at this station is \(~2.8\ m\). As sea level rises, the maximum \( u \) velocity decreased from 0.36 m s\(^{-1}\) in the present to 0.15 m s\(^{-1}\) in the Highest1 and 0.26 m s\(^{-1}\) in the Highest2 scenarios. It again increased to 0.36 m s\(^{-1}\) under the Highest3 simulation (Figures 4-10e, 4-10f, 4-10g, 4-10h). The \( u \) velocity was smallest in the Highest1 scenario because the highest flooding extent occurred under this condition and subsequently produced the largest frictional effects (Figure 4-5 and Table 4-3). Under the Highest3 simulation where marsh was accreted by 100% of RSLR, the \( u \) and \( v \) magnitudes were close to the present condition but with 2.9 hr phase lead which indicates that tidal waves propagated faster in the Highest3 scenario compared to the present condition. It occurred because RSLR deepened the channels and bayous and consequently increased the conveyance effects of the estuary. Under the present condition, the main \( x \)-momentum balance at station M2 and during both ebb and flood phases were between DP and FRIC. This is consistent with previous studies suggesting that in frictionally dominated estuaries the lowest order dynamics is characterized by a zero-inertia equation, i.e., a balance between bottom friction and pressure gradient (LeBlond, 1978; Friedrichs and Madsen, 1992; Huang et al., 2011). Consequently, in the present sea level, tidal wave propagation at this location can be properly described as a diffusion rather than a wave propagation (LeBlond, 1978). As sea level rises, the contribution of both FRIC and DP decreased and the role of ADV and DDT increased, thus tides became more propagational (Figures 4-10e, 4-10f, 4-10g, 4-10h).

In the upper estuary and under the present sea level, the \( v \) velocity was greater than the \( u \) velocity because M3 station is located in Bayou des Allemends which is oriented mainly in a north-south direction (Figures 4-10i and 4-11i). The \( v \) velocity was symmetric with equal maximum positive and negative values (0.08 m s\(^{-1}\)). As sea level rises, both positive and negative \( v \) velocities were increased. Under the Highest1 scenario, the maximum positive and negative \( v \) were 0.17 m s\(^{-1}\) and 0.10 m s\(^{-1}\) respectively. It may suggest that the flow in this bayou was flood dominated in the Highest1 scenario, but the ebb flow (negative \( v \)) lasts at its maximum status (0.10 m s\(^{-1}\)) for most of the ebb cycle (~10 hr). The momentum balance for M3 station was more complex than other stations. Under the present sea level condition, the main \( y \)-momentum balance was among ADV, DP and FRIC. As sea level rises, the contribution of ADV and DDT increased which indicates that the tide became more and more a wave phenomenon rather than diffusion of tidal signal. Comparing the tidal phase in three stations suggested that at the estuary mouth, tidal phase does not change much with sea level rise, while in the middle and upper estuary the tidal phase has obvious changes. This feature was also shown before for individual \( K_1 \) phases (Figure 4-7).

### 4.4 Discussion

Model simulations predict that tidal range will increase for all scenarios of sea level rise and marsh vertical accretion (Figures 4-6 and 4-7). Several previous studies have reported that if low-lying land is allowed to be inundated, tidal range will decrease in the estuary with sea level rise. For example, Lee et al., (2017) found that tidal range decreases in both Chesapeake Bay and Delaware Bay when low-lying land is allowed to become permanently inundated by higher sea level. Holleman and Stacey (2014) found a similar result in a modeling study of San Francisco Bay. In those estuaries which are much deeper than Barataria Basin, when low-lying land is allowed to flood, increased dissipation in newly inundated areas offsets reduced dissipation in
deeper water and therefore causes an overall reduction in the tidal range (Lee et al., 2017; Holleman and Stacey 2014; Pelling and Green 2013). In contrast, our modeling results indicated that even in the Highest1 scenario when no accretion was applied and extensive low-lying lands were flooded, tidal range increased all over the bay. Generally, RSLR increases the frictional effects by flooding of the low-lying lands and at the same time increases the conveyance effects through deepening of the existing channels and enhancing water exchange through newly flooded areas. However, if the estuary is deep, increased conveyance effect may be small or negligible compared to the increased frictional effect. In contrast, if the estuary is shallow like Barataria Basin, increased conveyance effect maybe larger than increased frictional effect. This is why model simulations predict that tidal range will increase even when extensive lands were flooded in Barataria Basin. Increased conveyance effects can also be inferred from $K_1$ phase differences between the mouth and the head of the estuary, which indicate that tidal waves traveled faster in all future RSLR scenarios compared to the present sea level conditions (Table 4-3).

The above reasoning can also explain why in the lower and the middle bay, the largest increase in tidal range occurred when the marsh area was assumed to keep pace with RSLR (less inundation). However, in the upper bay the largest increase in tidal range occurred when no accretion was assumed (higher inundation). This is because the lower and the middle bay are relatively deeper, and they have higher water exchange with the coastal ocean. In contrast, the upper bay is shallower and consists of extremely choked bayous and lakes. Therefore, flooding of the low-lying lands in the lower bay increases the frictional effects more than conveyance effects but in contrast in the upper bay, flooding of the low-lying lands increases the conveyance effects more than frictional effects. Comparing the $K_1$ phase difference between the mouth and the head of the estuary in different scenarios confirms this reasoning as it indicates that tidal waves traveled fastest in scenarios where no accretion was assumed (Table 4-3). It also explains why tidal amplification in the upper bay only occurred under the Lowest1, Medium1 and Highest1 scenarios (Figures 4-6, 4-7g and 4-7h). Tidal amplification is also reported in the modeling studies of tides on other estuaries. Lee et al., (2017) demonstrated that sea level rise induced tidal amplification in the upper part of Chesapeake Bay and Delaware Bay. Van Rijn (2011) found that in sufficiently long, deep, and converging estuaries, the amplifying effects dominate, and tidal amplitude increases toward the head of the estuary. Holleman and Stacey (2014) demonstrated that increased mean sea level, while preserving original shorelines, produces additional tidal amplification in San Francisco Bay. A detailed momentum balance analysis in this study indicated that sea level rise shifts tidal wave in Barataria Basin from a dissipative tidal regime to a progressive wave which is more susceptible to tidal amplification (Figures 4-10 and 4-11).

4.4.1 Limitation of the analysis

This research further demonstrates the importance of moving beyond bathtub modeling approaches by shifting the paradigm of RSLR assessments to approaches that account for the coastal dynamics of sea level rise (Passeri et al., 2015b). By more completely representing and examining the geophysical characteristics of the system we have further distinguished the nonlinearity of RSLR. However, predictions of this study still have limitations. Examples of such limitations include the accretion of the channels and bay bottom which were not considered. Even in simulations where the marshes were assumed to accrete at the same rate of RSLR, the
bottom did not have any accretion. However, in reality the bottom of the estuary can accrete if some sediment is imported into Barataria Bay, for example from the coastal ocean (Payandeh et al., 2020), from the GIWW (Mariotti et al. 2021), or from the David Pond diversion (Keogh et al., 2019). Furthermore, the model did not account for changes in marsh extent. Lateral retreat of the marshes can widen the channels, and thus affect tidal propagation. This is however difficult to predict because many marsh edges are now armored and protected. Furthermore, uncertainties associated with digital elevation models (DEMs) could affect the inundations predicted by the model and therefore affect the tidal response in the future. A higher resolution and more accurate bathymetry data in shallow waters of the northern Gulf of Mexico will benefit modeling coastal processes of the region in the future.

4.5 Conclusion

The results of this study quantify the combined effects of RSLR and marsh accretion on tidal dynamics in a tidally choked estuary. Under the present sea level condition, five major tidal choking areas were identified within the bay where tidal ranges were reduced sharply, and phase lags were developed. RSLR reduced tidal choking intensity and thus increased tidal range within the estuary. Despite the previous modeling studies in other estuaries suggesting that flooding of the low-lying land with sea level rise would increase tidal dissipation and thus reduce tidal ranges, here, it was shown that tidal range in an extremely choked tidal system like Barataria Basin will increase even when extensive lands are flooded. This is because the channel conveyance effect is larger than the frictional effect of the low-lying areas. The most interesting result of this study is that the change in tidal range varies along the estuary axis, and they are strongly dependent on the marsh vertical accretion. In the lower and the middle bay, the largest increase in tidal range (up to 16 cm) occurred when the marsh area was assumed to keep pace with RSLR. In the upper bay the largest increase in tidal range (up to 13 cm) occurred when no accretion was assumed. RSLR also induced amplification of tides at the head of the estuary. A detailed momentum balance analysis indicated that sea level rise shifts tidal wave from a dissipative tidal regime to a progressive wave which is more susceptible to tidal amplification. The positive feedback between RSLR and higher tidal ranges contributes to rapidly increasing inundation in the future. In the less optimistic condition, it was predicted that 75% of the domain will be permanently flooded while the remaining 25% will also be partially flooded. Given that inundations reported in this study are only tidally induced, the actual inundations would be larger if subtidal water levels are added to the RSLR models. After all, while this paper focused on the impact of RSLR and marsh accretion on tidal dynamics, it is important to note that other geomorphology changes could happen in the future. Changes in the bay bottom, the lateral marsh extent, and the barrier islands erosion. Therefore, additional research is required to comprehend the full impacts of RSLR and geomorphology changes on tidal dynamics in Barataria Basin.
CHAPTER 5. SUMMARY

In this dissertation, tidal and subtidal dynamics in Barataria Bay are investigated. The overall research objectives were:

1) Quantifying the relative importance of local and remote winds on subtidal water level and currents behavior in Barataria Bay during cold front season.

2) Exploring the seasonal variability of suspended sediment concentration in Barataria Bay, as the region transitions from a period of low river discharge and high cold front activity to a period of high river discharge and low cold front activity.

3) Investigating the combined effects of relative sea level rise and marsh accretion on tidal dynamics in Barataria Bay.

In order to address these research objectives a combination of field measurements and numerical modeling was used.

In chapter 2, to quantify the relative importance of local and remote wind effects on subtidal dynamics, three different methods were used: (i) statistical analysis of field observations, (ii) an analytical model, and (iii) a 2-D barotropic numerical model. Results indicated that water level variations in Barataria Bay were largely affected by cold fronts. The remote and local wind forcings were equally important at the bay mouth, however inside the bay the local forcing was dominant. The amplitudes of subtidal water levels associated with local winds were at least twice as large as those induced by remote winds, notably at the bay head. This finding differs from those found in the existing literature, notably in Breton Sound and Lake Pontchartrain, where remote wind dominates over local winds (Snedden et al., 2007; Huang & Li, 2017). These differences are attributed to the different geomorphological features of the estuaries. For instance, alongshore currents in the Northern Gulf of Mexico are intercepted by the Mississippi bird-foot delta and Barataria Bay is located on west side of the delta. Additionally, Barataria Bay is a Bar-Built estuary and is restricted on the south end by barrier beaches and Barrier islands. Therefore, cross-shelf Ekman transport is not fully established in this region, which further decreases co-oscillations between the shelf and the estuary. Given that the local effect is proportional to the onshore component of the wind and the remote effect is proportional to the alongshore component, remote and local winds operate nearly independently in Barataria Bay because the estuary is almost perpendicular to the coast (≈340°). This is not the case for Lake Pontchartrain, for example in which the two wind components act in concert since estuarine axis is nearly parallel to the coast. The east west orientation of the Louisiana coast also favors initial setup during the prefrontal phase of the cold fronts and subsequent setdown in frontal phase.

Better understanding of subtidal dynamics is instrumental for evaluating the effects of hydrologic restoration, and hydrodynamic consequences associated with proposed river diversion in Barataria Bay.

In chapter 3, the seasonal variability of the SSC in Barataria Bay was explored. Two rounds of observations were conducted. The winter deployment was characterized by low river discharge and high cold front activity and the spring deployment was characterized by high river discharge and low cold front activity. The SSC dynamics in Barataria Pass during the winter was mostly driven by resuspension in response to cold front passages. In addition to cold fronts, diurnal tides were also a controlling mechanism. During the spring, the average SSC was significantly higher than winter because of the strong offshore influence of the Mississippi River plume. Existence of freshwater on the estuarine mouth was found in long term monthly salinity
measurements along the estuary conducted by Turner et al., (2019). Further, MODIS satellite imagery during March 2018 confirmed the intrusion of the river plume though the Barataria pass. Cold fronts enhanced sediment export during winter. In contrast, cold fronts increased sediment import during the spring as they were associated with the higher offshore SSC coming from Mississippi River Plume.

In chapter 5, tidal change in response to the relative sea level rise and marsh accretion in Barataria Bay was investigated. Nine distinct simulation scenarios categorized under the major classes of lowest, medium, and highest relative sea level rise rates and different marsh accretion levels were implemented using a Finite Volume Community Ocean Model (FVCOM). Under the present sea level condition, five major tidal choking areas – where tidal range reduces sharply, and phase lags develop – were identified. Tidal choking intensity was reduced by relative sea level rise and thus tidal range were increased within the estuary. Contrary to previous modeling studies in other estuaries suggesting that flooding of the low-lying land with sea level rise would increase frictional effects and thus reduce tidal range, tidal range in an extremely choked tidal system like Barataria Basin increases even when extensive lands are flooded. This occurs because the channel conveyance effects are larger than the frictional effect of the low-lying areas. In the lower and the middle bay, the largest increase in tidal range occurred when the marsh area was assumed to keep pace with relative sea level rise. Under this condition, mean tidal range could increase by a maximum of 16 cm compared to the present sea level condition. However, in the upper bay the largest increase in tidal range occurred when no accretion was assumed. Under this condition, mean tidal range could increase by a maximum of 13 cm.

Relative sea level rise caused tidal amplification at the head of the estuary. Momentum balance analysis indicated that relative sea level rise shifts tidal regime from a dissipative tidal wave to a more progressive wave, which is more likely to be amplified. The positive feedback between relative sea level rise and higher tidal ranges contributes to rapidly increasing inundation in the future.

While in chapter 4, I focused primarily on tidal changes in response to relative sea level rise and marsh accretion, it is important to note that other geomorphology changes could occur in the future. Changes in the estuarine bottom and inlet areas, as well as the loss of barrier islands may significantly affect tidal dynamics in the future. Therefore, additional research is required to fully explore the potential impacts of relative sea level rise and changes in geomorphology on tidal dynamics in Barataria Bay.
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Suspended sediment dynamics in a deltaic estuary controlled by subtidal motion and offshore river plumes.

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VITA

Ali Reza Payandeh was born in Tehran, Iran, in 1987. He graduated high school from Abou Torab School in Qods, Iran. He then acquired his Bachelor of Sciences degree in Civil Engineering from University of Mohaghegh Ardabili in 2011. He started his master’s program in Coastal and Environmental Engineering in the College of the Environment, University of Tehran in 2012. He earned his Master of Science degree after completing a thesis working on the development of a carrying capacity model for nutrients to predict coastal eutrophication in estuaries. Subsequently, he gained extensive hands-on experience as a coastal engineer and modeler participating in a number of coastal engineering projects where he implemented 2-D and 3-D models to simulate coastal circulation, tsunami inundation and wave dynamics. He had two years of professional experience prior to enrolling a Ph.D. program at Louisiana State University. He became a candidate for the Doctor of Philosophy degree in 2020 and is expected to graduate in August 2021.