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PHYSICAL AND BIOLOGICAL FACTORS CONTROLLING THE FATE OF NITRATE IN A LOUISIANA COASTAL DELTAIC FLOODPLAIN

A Dissertation

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

In

The Department of Oceanography and Coastal Sciences

by

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B.S., University of California, Berkeley, 2011
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May 2020
For Henry and Peggy Lew
Acknowledgements

I would like to acknowledge my committee: Dr. Robert Twilley (chair), Dr. Clinton Willson, Dr. Ehab Meselhe, Dr. Laurel Larsen, Dr. Giulio Mariotti, and Dr. Yi-Jun Xu for their mentorship throughout this project.

This research was funded in part through the National Science Foundation Frontiers in Earth Surface Dynamics Grant (Award # 1135427), the National Science Foundation Coastal SEES Collaborative Research: (Award # 1427389), and Louisiana Sea Grant. My graduate funding was provided through the Coastal Flagship Assistantship and in part (Chapter 4) by The Water Institute of the Gulf under project award [“Multiple Tools for Determining the Fate of Nitrate in Coastal Deltaic Floodplains”].

The latter project was paid for in part with federal funding from the Department of the Treasury through the Louisiana Coastal Protection and Restoration Authority’s Center of Excellence Research Grants Program under the Resources and Ecosystems Sustainability, Tourist Opportunities, and Revived Economies of the Gulf Coast States Act of 2012 (RESTORE Act). The statements, findings, conclusions, and recommendations are those of the author(s) and do not necessarily reflect the views of the Department of the Treasury, CPRA or The Water Institute of the Gulf.

Sample analysis was completed through the Wetland Biogeochemistry Analytical Services Lab at Louisiana State University (Tommy Blanchard and Sara Gay) and field support and boat access were provided through the College of the Coast and Environment Field Support Shop (Mark Miller) and Dauterive Contractors.

In addition, I would like to acknowledge my research group, the Coastal Systems Ecology Lab, those who have assisted with field and lab work, and close friends: Edward
Castaneda-Moya, Azure Bevington, Leanna Heffner, Anika Aarons, Andre Rovai, Song Li, Annabeth McCall, Nicholas Olsen, Denise Poveda, Darian Madere, Allison Benelli, Madeline LeBlanc, Christina Chapman, Nadia Hamed, Alex McClellan, Jiaze Wang, Sara Gay, Tess Danielson, Kathleen Eubanks, Amanda Fontenot, and Zoe Shribman. A special thanks to Nicholas Olsen, my partner and the most dedicated field volunteer, who supported throughout graduate school.
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Abstract

The Mississippi River Delta is threatened by a growing pressure to support large human populations in the United States both with food production, navigation systems, and urban development in the Mississippi River Basin. Nitrate-nitrogen load in the Mississippi River, up to 100 Tg N yr\(^{-1}\) from agricultural and urban runoff, leads to phytoplankton blooms and hypoxia across the Louisiana continental shelf, creating dead zones of low dissolved oxygen threatening a significant commercial fishery. Along the coast and river corridors, floodplain ecosystems have the capacity to retain and remove nitrate. This dissertation explores the role of productive, actively growing coastal deltaic floodplains as nitrate removal hotspots in an active delta. A combination of field, laboratory, and numerical modeling tools are utilized in studying the physical and biological factors that control the nitrate removal potential of coastal deltaic floodplains. The overall impact of coastal deltaic floodplains in removing nitrate is contingent on connectivity or delivery of channel water to floodplains, which is spatially and temporally variable. Field studies indicate that denitrification removes nitrate on long-term scales and accounts for 28-71% of the nitrate fluxes across WLD floodplains. Water column assimilation, on the other hand, only temporarily removes nitrate as remineralization and decomposition recycles organic nitrogen back into bioavailable forms. Storage within sediment and vegetation compartments account for only 1-26% of nitrate fluxes, indicating slow transfer rates from surface water to the marsh sediment and vegetation. Longer water residence times within the floodplains of WLD facilitate greater water retention that promotes nitrate removal. However, residence time within floodplains reduces from 1.2 to 0.5 day as river discharge increases. Similarly, areas with the greatest denitrification rates and longest residence times represent a relatively small area of the active delta. Water quality models of WLD indicate only 2.4-4.1% of
nitrate loaded from Atchafalaya River is removed prior to export to Gulf of Mexico. Therefore, the overall contribution of WLD to long-term nitrate removal may be small considering the large nitrate load in this system. Results from this research highlight the need to consider both physical and biological conditions when designing restoration projects aimed at improving water quality along the coast of Louisiana.
Chapter 1. Review of Coastal Deltaic Floodplains

1.1. Introduction

With the invention of the Haber-Bosch process, production of nitrogen fertilizers increased from <10 Tg N yr\(^{-1}\) in 1950 to 100 Tg N yr\(^{-1}\) in 2000 and is expected to increase to 96-118 Tg N yr\(^{-1}\) by 2030 (Galloway et al., 2003). Although these fertilizers have increased food production to support the rapidly growing population in the United States, most scientists now recognize the large impact of this agricultural change on the environment (Erisman et al., 2013; Galloway et al., 2008; Vitousek et al., 1997). Rain and irrigation carry excess fertilizers to local streams and eventually to the Mississippi River, which drains 44% of the contiguous United States and carries large volumes of water, dissolved compounds, sediment, and other particulates. Agricultural, urban, and industrial runoff have led to increased loads of bioavailable nitrogen, primarily nitrate, to the Gulf of Mexico (GOM, Goolsby and Battaglin, 2001). As a key nutrient for many organisms, nitrate in the GOM stimulates phytoplankton populations creating algal blooms as the river plume mixes with coastal ocean. These bloom populations die and sink to bottom shelf waters where microbial decomposition can deplete dissolved oxygen in deeper water strata and produce hypoxic zones in water depths between 50 to 100 m in the Louisiana Bight. While hypoxia is a naturally occurring phenomenon in many stratified coastal waters, the spatial and temporal extent of hypoxia has increased significantly along the coast of Louisiana due to high nitrate concentrations in the Mississippi River and its distributaries (Rabalais et al., 2002). (Fig. 1.1).

Opportunities for nitrate removal exist throughout the Mississippi River watershed and continue to the delta of the Louisiana coast (Mitsch et al., 2005a, 2001). Alluvial floodplains of the river network and coastal deltaic floodplains and estuaries of the delta (Twilley et al., 2019)
have the capacity to retain and remove nitrogen. Thus, these wetlands provide ecosystem services by alleviating water quality problems downstream (Brinson et al., 2010; Racchetti et al., 2011).

Figure 1.1. May dissolved nitrate plus nitrite load to the Gulf of Mexico and area of mid-summer bottom water hypoxia (dissolve oxygen concentrations of less than 2 milligrams per liter) in the northern Gulf of Mexico. Source: United States Geological Survey

Human infrastructure has altered the delta cycle by channelizing the Mississippi River through the construction of levees causing accelerated rates of degradation to the Mississippi River Delta (Syvitski et al., 2009). While the Mississippi River once fed its floodplains with sediment and nutrients, channelization now prevents overbank flooding, disconnecting it from the floodplains. Instead, the main channel of the Mississippi River now carries sediment and nutrients to the continental shelf (Gagliano et al., 1981; Kesel, 1988)). The sediment falls out in deep water of continental shelf slope and canyons, with little deposition contributing to coastal morphology. Lacking sediment from the river, floodplain wetlands subside and erode due to
elevation deficit, leading to coastal land loss that has been the focus of coastal issues in Louisiana (Couvillion et al., 2011; CPRA, 2017).

To the west of Mississippi River Delta, active floodplains of the Atchafalaya Coastal Basin are connected to the Atchafalaya River, a distributary of the Mississippi River. The Atchafalaya Coastal Basin consists of deltaic floodplains in the proximal region (Atchafalaya Delta and Wax Lake Delta (WLD)) and deltaic estuaries in the distal region (Atchafalaya Bay and Fourleague Bay) (Twilley et al., 2019). Connectivity between the Atchafalaya River and adjacent floodplains allows exchange of nutrient-enriched riverine water and increased potential for nitrate removal. With minimal impact from navigational dredging and consistent monitoring since its development, WLD is an ideal case study for delta development and nitrate-removal potential in coastal Louisiana.

The U.S. Army Corps of Engineers dredged Wax Lake Outlet as a flood canal from Atchafalaya River to reduce the risk of flooding in Morgan City, Louisiana in 1942 (Fig. 1.4 a-b; Fisk, 1952; Latimer and Schweizer, 1951). This simulated stream capture and artificially began a new delta cycle at WLD (Roberts, 1997). Water and sediment were diverted through the outlet, delivering sediment to the river mouth. During a large flood in 1973, enough sediment was deposited for the first jet-plume deposit of WLD to remain exposed at mean low tide (Roberts, 1998). Three recent flood years (2008-2010) had an average annual discharge through the outlet of 109 km$^3$ yr$^{-1}$ with a 20.5 Mt yr$^{-1}$ sediment load (Fig. 1.2; Allison et al. 2012). With a sustained sediment supply and little impact from human disturbances, WLD continues to grow at approximately 1-2 km$^2$ yr$^{-1}$. 
Deltaic floodplains, as with alluvial floodplains, receive water, sediment, and nutrients from primary distributary channels through overbank flow and groundwater discharge (Fig. 1.3). In addition, secondary distributary channels can deliver water directly to the interior of the floodplain. Therefore, river stage impacts the hydrology of deltaic floodplains, moving water predominantly downstream. However, in coastal deltaic floodplains, marine forcings, such as tides and waves, also push water upstream and hourly water levels are highly influenced by the local tidal signature and wind patterns (Hanegan and Georgiou, 2014). During episodic events, such as hurricanes and cold fronts, water is forced further up into the delta and can play a large role in the hydrology and sediment transport of deltaic floodplains (Bevington et al., 2017).

The floodplains of river-dominated deltas, such as WLD, consist of arrowhead islands and interdistributary bays (Shaw et al., 2016). Islands are defined as the areas above mean low water (MLW), while the interdistributary bays are defined to include shallow areas (below MLW) within or just downstream of the islands. Deltaic islands often begin where mouth bars previously caused channel bifurcation. At the upstream portion of the island, natural levees
confine the flow and limit access of channel water into the island interiors. Further downstream, elevations along the edge of the island decrease and exchange between the channel and the floodplain increases. Across the island, elevation decreases towards the center of the island as the island transitions to an interdistributary bay (Johnson et al., 1985). This elevation gradient directs the flow of water and can lead to pooling of water in mid-elevation interior patches. Dye studies suggest that water residence time can be on the order of days in the interior of the island, which is much longer than the hourly residence times in the channels (Hiatt and Passalacqua, 2015a).

Figure 1.3. Conceptual model of a coastal deltaic floodplains with connectivity to the primary distributary channels of the delta driven by overbank flow, hyporheic flow, tides and cold fronts, and direct delivery from secondary distributary channels.

The geomorphologic features of coastal deltaic floodplains are classified into hydrogeomorphic (HGM) zones, defined by elevation relative to the local tidal datum (Fig. 1.4 b-d; Bevington et al. 2018). Supratidal floodplains occur along the levees where high elevations limit inundation. Intertidal and subtidal floodplains occur at mid-elevations (<MHW and >MLW) and low elevations (<MLW), respectively. Through delta formation, patterns of marsh age occur across coastal deltaic floodplains (Bevington and Twilley, 2018). Age and elevation
are key characteristics in understanding both physical and biological spatial patterns of deltaic floodplains, including soil development and vegetation structure.

Figure 1.4. (a) The study site, Wax Lake Delta, located in Atchafalaya Bay in southeastern Louisiana. (b) Hydrogeomorphic zone classification of Wax Lake Delta using USGS National Elevation Dataset Digital Elevation Models derived from 2012 Lidar data. (c) Conceptual diagram of exchange between primary distributary channels and floodplains on Mike Island. (d) Aerial view of Mike Island, highlighting hydrogeomorphic zones.

Vegetation also affects deltaic island hydrology by providing significant resistance to water flow, which can enhance mineral sedimentation. The effect of vegetation depends on the height to water depth ratio, diameter, flexibility, and water velocity (Kadlec, 1990; Temmerman et al., 2005a). Dominant species vary throughout WLD, controlled mostly by elevation and inundation. *Salix nigra* establish along natural levees with reduced inundation. As elevations
decrease, the dominant species changes: *Colocasia esculenta* (mean elevation of 0.57m), *Polygonum spp.* (mean elevation of 0.47m), *Nelumbo lutea* (mean elevation of 0.21m), SAVs (mean elevation of 0.15m), and *Potomogeton nodosus* (mean elevation of 0.07m) (Fig. 1.5; Carle et al., 2015).

![Diagram of vegetation species at various elevations within Wax Lake Delta](image)

**Figure 1.5.** Dominant vegetation species at various elevations within Wax Lake Delta. *Source: Carle et al., 2013*

Geophysical and ecological factors contribute to the unique hydrology of coastal deltaic floodplain wetlands, creating environments of longer residence times where important biogeochemical processes occur. Hydrology controls the connectivity between a river and adjacent floodplains. This connectivity determines the temporal and spatial extent of sediment and nutrient delivery to floodplain wetlands. Previous field studies by Castaneda et al. (unpublished data) demonstrate the impact of hydrogeomorphic zones on flow patterns and water quality parameters at WLD. Their survey of deltaic islands highlights trends in velocity, temperature, and surface water nitrate concentration. Lower nitrate concentrations occur within the intertidal floodplains, which are characterized by mid-elevations, slower velocities, and warmer water temperatures. In comparison, faster velocities, cooler temperatures, and higher nitrate concentrations characterize subtidal floodplains. These trends are most apparent in March.
and April when river water temperatures are low. As river temperatures warm, differences in temperature and nitrate concentrations throughout the island become less distinct.

The nitrogen cycle of a floodplain wetland includes several processes, involving inorganic nitrogen (ammonium, nitrite, nitrate, dinitrogen gas) and organic nitrogen. nitrate, ammonium, dinitrogen gas, and organic nitrogen. Many of these processes are important for reducing the concentration of bioavailable nitrogen that is transported by the Mississippi River to the GOM, including burial, plant uptake, denitrification, and dissimilatory nitrate reduction to ammonia (DNRA) in wetlands. Plant uptake and DNRA are only short-term pathways because they do not remove nitrogen from the system. Instead, some nitrate used by plants is recycled at the end of the grow season when plants die. Similarly, DNRA transforms nitrogen from one bioavailable form to another, producing ammonium that also feeds phytoplankton. Burial and denitrification, on the other hand, are long-term pathways, effectively removing nitrate from channels flowing into coastal ocean. Denitrification is the conversion of nitrate to dinitrogen gas as part of microbial respiration. In the absence of oxygen, nitrate is a favorable electron acceptor for several heterotrophic bacteria and is reduced as organic matter is oxidized.

Physical and biological factors, most notably organic matter content, concentration of NO$_3$ in river water, inundation frequency and duration, and temperature, control denitrification rates. At WLD, denitrification increases with sediment organic matter content, which increases with marsh age (Fig. 1.6; Henry and Twilley, 2014). However, lower ambient nitrate concentrations produce lower denitrification rates and greater nitrogen fixation rates. As an anaerobic process, denitrification occurs in frequently flooded soils, highlighting the importance of inundation on denitrification rates. Elevation and river stage are two factors controlling frequency and duration of flooding. Even small changes in deltaic island topography can create
variations in inundation. Finally, as a microbially-mediated process, denitrification increases with temperature, with optimal temperatures between 30-60°C. Overall, denitrification rates vary spatially and temporally, producing biogeochemical hotspots, areas of increased nitrate removal (McClain et al., 2003; Vidon et al., 2010). Studying the nitrogen dynamics WLD can help to answer questions about the fate of nitrate in coastal deltaic floodplains.

Figure 1.6. Relationship between (a) organic matter content and island age and (b) organic matter content and N$_2$ fluxes. Source: Henry and Twilley, 2014

Field measurements and laboratory experiments are important for observing trends and determining biogeochemical rates that may process nitrate from rivers prior to export to coastal ocean. However, numerical models are essential to answering larger scale questions. Numerical models can be used to simulate biogeochemical processes, but must be coupled with an appropriate hydrodynamic model (Hopkinson and Day, 1980; Mitsch and Reeder, 1991). Modeling the complex flow paths through an inter distributary deltaic island could aid in understanding the movement and transformation of NO$_3$ and other nutrients within this terrestrial interface with coastal ocean. While difficulties in modeling biogeochemical hotspots have been
previously presented (Groffman et al., 2009), a high-resolution water quality model (Delft3D – Water Quality) combined with field data and rigorous laboratory experiments could help to define the fate of nitrate. A plethora of hydrodynamic-biogeochemical models have been utilized for estuarine and terrestrial ecosystems (Grégoire and Friedrich, 2004; Smits and van Beek, 2013) and can be used to as a starting point for modeling the fate of nitrate in deltaic wetlands.

1.2. Objectives

The following chapters combine field, laboratory, and numerical modeling experiments to explore the role of coastal deltaic floodplains in nitrate removal potential and develop tools for studying the coupling of hydrodynamics and biogeochemistry of coastal wetlands. Through this dissertation, the following objectives are met:

- Quantify the exchange between primary distributary channels and deltaic floodplains and the average residence time of water within floodplains of a young, prograding river-dominated delta.
- Identify spatial patterns in vegetation structure and soil parameters across the floodplains of a prograding delta and develop methods for assigning them across a larger spatial scale.
- Quantify nitrate removal potential and a general nitrogen budget of an individual island of Wax Lake Delta.
- Apply nitrate uptake rates within a hydrodynamic numerical model to explore the nitrate removal capacity of Wax Lake Delta.
Chapter 2. Simulating Connectivity and Water Exchange within a Coastal Deltaic Floodplain of the Mississippi River Delta

2.1. Introduction

Coastal deltaic floodplains are distinct geomorphic units that form at the mouth of major rivers where jet-plume deposits develop (Islam, 2016; Syvitski et al., 2012; Twilley et al., 2019; Wellner et al., 2005). Similar to alluvial floodplains, coastal deltaic floodplains are flooded by primary distributary channels when river stage reaches the elevation of natural levees, transporting sediment along with dissolved and particulate nutrients from channels to floodplain ecosystems (Amoros and Bornette, 2002; Asselman and Middelkoop, 1995; Noe and Hupp, 2009). These floodplains are transition zones between terrestrial and aquatic ecosystems where wetlands form as land emerges from subaqueous environments (Twilley et al., 2019). Wetlands colonizing deltaic floodplains have been associated with several ecosystem services such as trapping sediment (Li and Yang, 2009; Nardin and Edmonds, 2014), reducing storm surge (Barbier et al., 2013; Wamsley et al., 2010), and processing riverine nutrients (Breaux et al., 1995; Mitsch et al., 2005b, 2001; Perez et al., 2011) making deltaic floodplains significant coastal landscapes at the edge of continents. However, several of these ecosystem services, such as sediment deposition and nutrient reduction, depend on the hydrological connectivity, or the amount of water exchanged between channel and floodplain ecosystems prior to discharge to the coastal ocean (Amoros and Bornette, 2002; Covino, 2017; Junk et al., 1989).

In addition, water residence time and water temperature within deltaic floodplain wetlands plays a significant role in the effectiveness of coastal deltas in processing nutrients (Hiatt et al., 2018; Hiatt and Passalacqua, 2015b; Twilley et al., 2019). Increased residence time extends time for nutrient uptake by phytoplankton and nitrate reduction through denitrification, a form of microbial respiration (Dettmann, 2001; Kaushal et al., 2008; Klocker et al., 2009). Thus,
in many lakes, rivers, and estuaries, nitrogen retention and export can be predicted using residence time (Dettmann, 2001; Nixon et al., 1996). Water age, a complement to water residence time, the amount of time elapsed since a parcel of water entered the system from a particular source can be modeled using particle tracers (Deleersnijder et al., 2001; Delhez et al., 1999; Shen and Haas, 2004; Viero and DeFina, 2016). Similarly, carbon, nitrogen, and phosphorus cycling are impacted by water temperature, particularly microbially-mediated processes (Bachand and Horne, 1999; Kadlec and Reddy, 2001; Rahmeyer and Werth, 1996; Stanford et al., 1975). Water temperatures within floodplains are often warmer than adjacent channels, depending on connectivity, water depth, water residence time, and climate. Therefore, connectivity, residence time, and water temperature are key factors controlling ecological processes and feedbacks of coastal deltaic systems. The connectivity between primary channels and floodplains as well as residence time and water age depend on the morphology of the floodplain, specifically surface topography and vegetation, and the hydrologic forcings of the delta system.

Deltaic floodplains receive water, sediment, and nutrients from primary distributary channels through river discharge into secondary distributary channels, overbank flow of natural levees, and hyporheic exchange (Heiler et al., 1995; Junk et al., 1989; Tockner et al., 1999). Therefore, river stage has a large impact on deltaic floodplain hydrology, moving water predominantly downstream to the coastal ocean. However, in coastal deltaic floodplains, marine forcings, such as tides and waves, also deliver water by moving coastal waters upstream (Everett et al., 2019; Li et al., 2011; Stern et al., 1986). Water levels are influenced hourly by the local tidal signature and wind patterns (Hanegan and Georgiou, 2014). Other episodic events, such as hurricanes and cold fronts, present special instances where tides and waves are forced further
inland into the delta. These events can play a significant role in the connectivity and sediment transport of coastal deltaic floodplains (Bevington et al., 2017; Hiatt et al., 2019; Roberts et al., 2015; Xing et al., n.d.). Resistance to flow derives from topography and wetland vegetation creating complex surface water exchange between channels and deltaic floodplains (Wright et al., 2018). Vegetation structure, density, rigidity, and submergence all contribute to the impact of vegetation on water flow through wetlands (Järvelä, 2002; Nepf, 1999). Together, river, tide, wind, and wave energy as well as physical features of the floodplain (i.e., geomorphology, size, stream channel, etc.) control the connectivity of water with wetlands on hourly, daily, and seasonal cycles, creating complex multi-directional flows that are distinct from alluvial floodplain connectivity (Anderson and Lockaby, 2012; Hiatt and Passalacqua, 2015b; Warne et al., 2002). Seasonal variation in vegetation and river stage, as well as tides, wind, and storm events, can shift the balance between flow and resistance, adding further complexity to coastal deltaic floodplain hydrology.

The connectivity of coastal deltaic floodplains, in contrast to alluvial floodplains, are driven by coastal forcings along with riverine forcings that control water stage and residence time (Hiatt et al., 2019; Hiatt and Passalacqua, 2015b). The floodplains of river-dominated deltas consist of arrowhead islands, areas above mean low water (MLW), and interdistributary bays, shallow areas below MLW within the islands, with deeper troughs down the island center (Ma et al., 2018; Shaw et al., 2016). Deltaic islands often begin where mouth bars previously caused channel bifurcation (Esposito et al., 2013). At the upstream portion of the island, natural levees confine the flow and limit access of channel water into the island interiors. Further downstream, island edge elevations decrease and exchange between the channel and the floodplain increases.
Across the island, elevation decreases towards the center of the island as the island transitions to an interdistributary bay (Johnson et al., 1985).

These geomorphic features of coastal deltaic floodplains are further characterized into hydrogeomorphic (HGM) zones (Bevington and Twilley, 2018), which are defined by elevation relative to a local tidal datum. Supratidal floodplains occur along levees where high elevations limit inundation. Intertidal and subtidal floodplains occur at mid-elevations (<MHW and >MLW) and low elevations (<MLW), respectively. Island chronosequence, together with elevation, is important in understanding physical and biological spatial patterns of deltaic floodplains, including soil development and vegetation structure (Bevington and Twilley, 2018). Patterns in biogeochemistry, such as benthic nutrient fluxes, vary with chronosequence depending on soil organic matter concentration that increases from subtidal to supratidal hydrogeomorphic zones (Henry and Twilley, 2014, Li et al, in review). These spatial patterns together with estimates of inundation duration and area of each HGM zone were used to estimate the capacity of coastal deltaic floodplains to remove nitrate at the mouth of the Atchafalaya River (Li et al., in review). However, these types of biogeochemical estimates of system capacity are based on simple considerations of flooding frequency and water residence time, which together with nutrient flux rates, determine nutrient processing rates in these tidal freshwater wetlands. Simulation models that can account for the complexity of connectivity between primary channels and floodplains will resolve better estimates of nutrient processing in coastal deltaic floodplains.

The complex hydrodynamics of coastal deltaic floodplains must be more clearly defined to understand how riverine nutrients are processed prior to export to the coastal ocean. This project aims to study the hydrologic interaction between delta channels and floodplains by
incorporating river, tide and wind forcings into a numerical Delft3D model. The main objectives are to: (1) compare hydrodynamics among HGM zones and the impact on water age and temperature; (2) quantify the hydrologic exchange and connectivity between channels and floodplains; and (3) test the impact of increased bed roughness (a proxy for vegetation density) and decreased tidal signal on connectivity and residence time. We hypothesize that water residence time and water temperature will increase at higher elevation HGM. We also hypothesize that exchange and connectivity will increase from the delta apex to the delta front and in response to increasing river stage during seasonal pulse flooding. Finally, we expect wetland vegetation density will decrease connectivity, but will increase residence time, while tides will increase connectivity and decrease residence time.

2.2. Methods

2.2.1. Site Description

Our study area, Wax Lake Delta (WLD), is a young (<40 years) and actively prograding delta located within the Atchafalaya Bay, in southeastern Louisiana (latitude 29 °N; Fig. 2.1a-b). WLD is a direct consequence of an U.S. Army Corps of Engineers project in 1942 that dredged the Wax Lake Outlet to reduce the risk of flooding in Morgan City, LA (Fisk, 1952; Latimer and Schweizer, 1951). Water and sediment were diverted through the outlet, stimulating stream capture and a new delta cycle. WLD became subaerial in 1973, after an extreme flood created a large sediment pulse through the Wax Lake Outlet (Roberts, 1998).

Three recent flood years (2008-2010) had an average annual discharge through the outlet of 109 km$^3$ yr$^{-1}$ with a 20.5 Mt yr$^{-1}$ sediment load (Allison et al., 2012). Eighteen percent of the sediment load is sand, while the delta sediment is 67% sand, which suggests that sand is preferentially deposited (Kim et al., 2009; Roberts, 1998). As more mineral sediment is delivered
and organic matter accumulates in the soil, WLD continues to grow at an estimated rate of 1-3 km\(^2\) yr\(^{-1}\) (Kim et al., 2009).

2.2.2. Model Description

The Delft3D modeling suite was chosen for this study because of its applicability to shallow water ecosystems and the option to couple hydrodynamics, heat flux, water quality, and sediment processes (Deltares, 2016). While this study focused on hydrodynamics and heat flux, future work will incorporate nutrient processes that can be modeled using the Delft3D-Water Quality module. As discussed above, capturing the hydrodynamics of deltaic floodplain wetlands was essential to modeling nutrient processes in these complex ecosystems.

Delft3D-FLOW solves the shallow water Navier Stokes equations using a finite difference method

\[
\left( \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} \right) = - \left( \frac{1}{\rho_0} \right) \frac{\partial p}{\partial x} + \nu \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial z^2} \right) - F_x \tag{Eq. 2.1}
\]

\[
\left( \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} \right) = - \left( \frac{1}{\rho_0} \right) \frac{\partial p}{\partial y} + \nu \left( \frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2} + \frac{\partial^2 v}{\partial z^2} \right) - F_y \tag{Eq. 2.2}
\]

\[
\left( \frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} \right) = - \left( \frac{1}{\rho_0} \right) \frac{\partial p}{\partial z} + \nu \left( \frac{\partial^2 w}{\partial x^2} + \frac{\partial^2 w}{\partial y^2} + \frac{\partial^2 w}{\partial z^2} \right) - F_z - \frac{\rho}{\rho_0} g z \tag{Eq. 2.3}
\]

where \(\rho_0\) = density; \(u, v,\) and \(w\) = velocity in the \(x, y\) and \(z\) directions; \(p\) = pressure; \(F_x, F_y, F_z\) = Reynolds stresses; \(g\) = gravitation acceleration; \(\nu\) = eddy viscosity.

Delft3D-FLOW also includes an ocean heat flux model, introduced by Gill (1982) and Lane (1989), which is used to model heat exchange and simulate spatial and temporal changes in water temperature. Total heat flux was calculated as

\[
Q_{tot} = Q_{sn} + Q_{an} - Q_{br} - Q_{ev} - Q_{co} \tag{Eq. 2.4}
\]
where $Q_{sn} = \text{net incident solar radiation (short wave)}$; $Q_{an} = \text{net incident atmospheric radiation (long wave)}$; $Q_{br} = \text{back radiation (long wave)}$; $Q_{ev} = \text{evaporative heat flux (latent heat)}$; $Q_{co} = \text{convective heat flux (sensible heat)}$.

The ocean heat flux model calculates clear sky solar radiation based on latitude, longitude and local time. For this model, the user must provide fractional cloud cover (%) for calculating short and long wave radiation, air temperature (°C) for long wave radiation, and relative humidity (%) for evaporative heat flux.

Delft3D-Water Quality (Delft3D-WAQ) is another module within the Delft3D modeling suite, aimed at simulating mass transfer of a variety of substances. Delft3D-WAQ can be coupled with the hydrodynamic output from Delft3D-FLOW with options for both spatial aggregation and longer time steps. Delft3D-WAQ solves the advection-diffusion equations, along with specific reaction processes defined by partial differential equations. Unlike Delft3D-FLOW, Delft3D-WAQ uses a finite volume method, which evaluates mass exchange across the surface of a volume. Delft3D-WAQ solves the advection diffusion equations:

$$M_i^{t+\Delta} = M_i^t + \Delta t \left( \frac{\Delta M}{\Delta t} \right)_T + \Delta t \left( \frac{\Delta M}{\Delta t} \right)_P + \Delta t \left( \frac{\Delta M}{\Delta t} \right)_S$$

Eq. 2.5

where $M_i^t = \text{mass of substance } i \text{ at time } t$; $\Delta t = \text{time step}$; $\Delta M = \text{changes in mass}; T_r = \text{transport}; P = \text{physical, biological, and chemical processes}; S = \text{sources}$.

### 2.2.3. Model Domain and Bed Elevation

The spatial extent of the model in this study included WLD and surrounding areas (Fig. 2.1b). The domain and the grid of the model were modified from Khadka (2013)(Khadka, 2013)(Khadka, 2013)(Khadka, 2013)(Khadka, 2013)(Khadka, 2013). For the purposes of this study, aimed at understanding flow characteristics of deltaic floodplain wetlands, rather than channels, the grid was locally refined to approximately
22 m x 16 m around Mike Island, the site used for calibration. The final grid has a total of 233,553 nodes.

Figure 2.1. (a) Delft3D model domain showing the hydrogeomorphic zones of Wax Lake Delta. (b) Conceptual diagram with Mike Island for reference indicating pathways of exchange between channels and floodplains.
Topography and bathymetry data were provided by Sadid et al. (personal communication), who compiled digital elevation models (DEM) of the delta from Shaw et al. (2016) and of the surrounding areas from USGS (1998). Shaw et al. (2016) constructed the DEM for the delta using multibeam bathymetry of the channels (recorded in 2007, 2009, and 2013), single beam bathymetry of the delta front (2015), and LiDAR of the subaerial marshes (2015). For more information on this dataset, see Shaw et al. (2016). Unfortunately, this DEM did not capture small secondary distributary channels, such as the one near the apex of Mike Island. As this channel is expected to be critical to the flow patterns on Mike Island, we surveyed the channel in March 2016 and incorporated it into our model DEM. A SonTek acoustic Doppler Current Profile (ADCP) measured the depth and discharge of the channel along six transects of the channel. Sediment bed elevations relative to NAVD88 were extracted by correlating water depths measured by the ADCP with water levels at the nearest tidal station in the Atchafalaya River Delta during the time of the survey.

2.2.4. Boundary Conditions

Publicly available field data were downloaded for the study period from January through June 2015 for model boundary conditions (Appendix A Fig. A.1, Table A.1). This time period represents a transition both in temperature and river stage, capturing important hydrodynamic and ecological processes within WLD. The upstream boundary conditions are defined within the Wax Lake Outlet by time series data from the USGS river station at Calumet (https://waterdata.usgs.gov/nwis/uv?site_no=07381590 Appendix A. Fig. A.2a). The downstream boundary conditions are defined along the southern edge of the domain by time series data from the NOAA tide station at Amerada Pass (https://tidesandcurrents.noaa.gov/stationhome.html?id=8764227 Appendix A. Fig. A.2b). The
The lower left and right edges of the domain are not defined as open boundaries as justified by Shaw et al. (2016). To test the appropriateness of this simplification, simulations with one downstream open boundary and with three downstream open boundaries were compared in this study and no significant impact was found on the hydrodynamics within the islands (data not shown).

Due to limited water temperature data in the WLO, upstream boundary conditions for the heat-flux model were set using hourly water temperature data from the Atchafalaya River at Morgan City, LA (https://waterdata.usgs.gov/usa/nwis/uv?site_no=07381600 Appendix A. Fig. A.2c). Downstream boundary conditions were set using data from Amerada Pass station (Appendix A. Fig. A.2d). The ocean heat-flux model required relative humidity, cloud cover, and air temperature to determine total heat flux from the free surface. Time series data of these parameters are available from the Patterson Airport and Salt Point (Appendix A. Fig. A.2e-g). While Patterson Airport is 30 km inland from WLD, differences in these environmental parameters are expected to be minimal and continuous data are not available at any closer locations.

2.2.5. Physical Parameters

Wind produces drag along the water surface causing surface water velocities to change that can propagate into the water column, especially in shallow systems. $C_d$, the wind drag coefficient, can increase linearly based on wind speed, $U_{10}$. For initial model setup, a piece-wise linear function between default $C_d$ values of 0.000663 at 0 m s$^{-1}$ and 0.0072 at 100 m s$^{-1}$ was used to determine $C_d$. Wind speed and direction time series (6-minute interval) data from the Amerada Pass tide station, located approximately 10m above the water surface, were applied uniformly across the entire domain (Appendix A. Fig. A.2h).
Hydrodynamic forces that connect water with deltaic floodplains such as river, tides, wind, and waves can be modified with resistance by sediment bed and vegetation. Together these physical and biological factors create distinct flow patterns within a deltaic floodplain. The effect of wetlands depends on vegetation height to water depth ratio, stem diameter, flexibility, and water velocity (Kadlec, 1990; Temmerman et al., 2005a). The effect of sediment bed roughness depends on topography and sediment particle size.

Sediment bed and vegetation create resistance to flow producing distinct flow patterns within a deltaic floodplain. The effect of vegetation depends on vegetation height to water depth ratio, stem diameter and flexibility, and water velocity (Kadlec, 1990; Temmerman et al., 2005a). The effect of sediment bed roughness depends on topography and sediment particle size. For simplicity, resistance due to vegetation and bed roughness were considered together in this model through a Manning $n$ coefficient. Initial $n$ values were assigned to the model domain based on the hydrogeomorphic zones outlined above. Vegetation type and structure have been shown to vary between these zones, likely a result of the effect that inundation frequency and duration have on vegetation community composition in floodplains systems (Blom et al., 1994; Carle et al., 2015; Harris, 1987). Floodplains, distinguished from channels using an elevation threshold of -0.75 m NAVD88, were classified into the three hydrogeomorphic zones, subtidal, intertidal, and supratidal floodplains by MLW (-0.04 m NAVD88) and MHW (0.30 m NAVD88) (Fig 1 b-d). Classifying the delta into hydrogeomorphic zones was supported by both field observations and surveys of vegetation density under different hydrologic conditions.

For this analysis, the study period was divided into three 2-month sections: January – February, March – April, and May - June. Colder temperatures, lower river stage, and sparse vegetation characterized the January – February period. From March – April, river stage
increased as the Atchafalaya River floods, while air and water temperatures rose, and the growing season began. Throughout May and June, river stage remained high, while warmer temperatures and longer daylength increased plant production (although peak biomass didn’t occur until August or September). Due to these seasonal differences, $n$ values for the subtidal and intertidal floodplains were changed and calibrated for each 2-month period.

2.2.6. Field Data and Calibration

A system of monitoring platforms within Mike Island, the Delta Dynamics Observatory (unpublished data), were equipped with instruments for measuring water quality parameters associated with river and marine forcings (unpublished data). These platforms were installed in 2014 and used for monitoring both short- and long-term trends that could support modeling efforts. We deployed six platforms, three along a longitudinal gradient from upstream to downstream locations in the interdistributary bay of Mike Island (Appendix A. Fig. A.3). These locations were designed to measure water behavior as a function of forcings from river floods, fronts and tides within different hydrogeomorphic zones. Each platform was equipped with sensors to measure water depth (pressure transducer, Campbell Scientific CS-451), temperature and conductivity (Campbell Scientific CS-547A), turbidity (OBS-500, Campbell Scientific), surface water nitrate concentrations (SUNA V2, chemical-free sensor, Satlantic), and water velocity and direction (Argonaut-ADV, Sontek). Data from all sensors were recorded at one-hour intervals and stored on a datalogger (Campbell Scientific CR-1000) in each platform.

Four of the platforms were used for model calibration and validation. Platform Mike Int-1 and Mike Int-2 were located within the intertidal floodplains and platforms Mike Sub-2 and Mike Sub-3 were located within the subtidal floodplains (Fig 2.1c). Gaps in data during January - February indicate low water levels (<18cm) when the instruments were not submerged. Data
from January – June 2015 were used to calibrate the hydrodynamic and heat-flux models using two variables: water level and water temperature. An initial sensitivity analysis was done to determine which of the parameters detailed above had the greatest influence on model output and should be used in calibration. In this analysis, each parameter was changed by +/- 25% and the average percent change in water depth, velocity magnitude, and water temperature at Mike Int-1, Mike Int-2, Mike Sub-2, and Mike Sub-3 were calculated as

\[
\text{Average } \% \text{ Change} = \frac{X_{x+25\%} - X_0}{X_0} \quad \text{Eq. 2.6}
\]

Following the sensitivity analysis, minor adjustments were made to the DEM, including a decrease in depths of the secondary channel on Mike Island (+0.65 m) and an increase in elevation of the levee at the northern end of the island (+0.5 m). In addition, upstream flow was reduced by 10% to further decrease the unexpected overbank flow. Results from the sensitivity analysis suggest that horizontal eddy viscosity, wind drag coefficient, and bed roughness had the greatest influence on model output. These parameters were used in full calibration. Sensitivity analysis indicated model response to the horizontal eddy diffusivity, Dalton number, and Stanton number was less than 3% for all output variables and thus initial values were used for the final model.

**2.2.7. Connectivity and Residence Time**

Exchange occurs when surface water crosses the boundary between channel and floodplain and connectivity describes exchange as a percentage relative to the discharge in the primary channels. Hyporheic exchange was not considered due to the limitations of the model. Exchange and connectivity were calculated during post-processing, using output maps from the model, which were stored every 2 hours. Three methods were used to calculate exchange and exclusively consider the major islands and channels of the delta (Fig 2.3a). In the first method,
we quantified flow into and out of floodplains by calculating the flow (m$^3$ s$^{-1}$) across edge cells. Edge cells were identified as floodplain cells with channel cells immediately adjacent, where cells were classified by an elevation threshold of -0.75 m NAVD88n (Fig. 2.3b). For the edge method, net exchange was calculated as

$$\text{Exchange}_{\text{edges}} = Q_{F_{\text{in}}} - Q_{F_{\text{out}}}$$

Eq. 2.7

where $Q_{F_{\text{in}}}$ is flow into floodplains across floodplain edges and $Q_{F_{\text{out}}}$ if flow out of floodplains across floodplain edges, determined from model output every two hours.

The second method was derived from the traditional alluvial calculation of fractional exchange. Exchange was calculated as the difference in channel discharge at the upstream and downstream boundaries of a channel section, which represents the amount of water leaving the channels and entering the floodplains.

$$\text{Exchange}_{\text{FE}} = Q_{C_{\text{upstream}}} - Q_{C_{\text{downstream}}}$$

Eq. 2.8

where $Q_{C_{\text{upstream}}}$ and $Q_{C_{\text{downstream}}}$ are discharges within the primary channels at an upstream and downstream boundary.

The third method evaluated exchange volumetrically, as the change in floodplain water volume over time. Water volume was calculated as the product of cell area and water depth. As the model did not include precipitation or evaporation, the only addition or removal of water to the delta was through surface water flow. Therefore, the change in volume within the floodplains represented exchange between channels and floodplains. For the volumetric method, exchange was calculated as the change in water volume within the floodplains over time:

$$\text{Exchange}_{\text{volumetric}} = \frac{\Delta V_f}{\Delta t}$$

Eq. 2.9

where $\Delta V_f$ is the change in floodplain water volume ($\Delta t$ime) is 2 hours.
Connectivity, which represents the percent of channel water flowing into the floodplains, can be estimated using the exchange rates from all three methods. However, the fractional exchange and volumetric methods only analyzed net-exchange at each time step, which incorporated both flow into and out of the floodplains. The edge method, on the other hand, separated flow in and out of the floodplains and more accurately represented the total delivery of water. Therefore, connectivity was calculated using the edge method as

\[ \text{Connectivity} = \frac{Q_{F_{in}}}{Q_{C_{in}} + Q_{F_{out}}} \]  

Eq. 2.10

where \( Q_{C_{in}} \) and \( Q_{F_{out}} \) are flows into channels (grey area in Fig. 2.2) across domain boundaries and floodplain edges, respectively, which together sum to the total channel flow (blue arrows in Fig. 2.2).

![Conceptual diagram for calculating connectivity using the edge method.](image-url)
We expected exchange and connectivity to vary from the delta apex to the delta front because of changes in the number of access points for exchange and the elevation of island edges. Thus, in addition to quantifying exchange for the entire delta, we also quantified exchange along a transect from the delta apex to the delta front. We divided the delta into eleven radial sections, approximately every 1000 m (Fig. 2.3c), considered each section as a separate control volume, and compared exchange and connectivity along this longitudinal axis. In addition, we utilized the edge method to test differences in exchange and connectivity among islands.

Figure 2.3. a) Delta area used in model post-processing to include all major islands, b) delineation of floodplain edges, and c) radial sections for upstream to downstream analysis

For exchange and connectivity analyses, we compared all three methods (edges, fractional exchange, and volumetric) to identify strengths and weaknesses of each method. To highlight areas of high and low agreement of channel-floodplain exchange, we analyzed a two-week window from February 1-14, 2015 (Fig. 2.4). Within the first seven radial sections where channel-floodplain exchange was small, there were relatively large discrepancies between each method (up to 200m$^3$s$^{-1}$). In the 7th and 8th radial sections (Fig. 2.4a-b), the edge method indicated a net flow of water from the channels into the floodplains, while the fractional exchange method and volumetric method had no net exchange and compare well with each
other. Further downstream, in the 8\textsuperscript{th}, 9\textsuperscript{th}, and 10\textsuperscript{th} radial sections, the edge method and volumetric method were more similar, but the fractional exchange method diverged, with much greater positive and negative extremes than the other two methods.

Figure 2.4. Comparing the edge method (blue), the fractional exchange method (black), and the volumetric method (red) for net channel-floodplain exchange.

Overall, the fractional exchange method was not appropriate to use for the entire delta (Fig. 2.4c) as it is expected that all water that enters at the delta apex (1\textsuperscript{st} radial section) will exit at the delta front (11\textsuperscript{th} radial section). Thus, accounting for the time lag due to river flow and tides, primary channel discharge at the delta apex and delta front should be equal. Alternatively, the volumetric method was not ideal for individual radial sections because this method cannot account for the exchange between the floodplains of adjacent radial sections. Considering these
limitations, we used the edge method and the volumetric method in delta-wide analyses and the edge method and the fractional exchange method when assessing delta apex to delta front trends.

Water age was calculated using Delft3D-WAQ, which simulates the movement of conservative and decayable tracers, released continuously at the upstream boundary of the model (Deltares, 2014). Both tracers move with the flow of water, but while the conservative tracer concentration is only controlled by transport, the decayable tracer concentration depends on a decay constant of 0.1 day$^{-1}$. Water age is calculated by Delft3D-WAQ as

$$age_{Tr_i} = \frac{\ln(\frac{dT_{Tr_i}}{cTr_i})}{RcDecTr_i} \quad \text{Eq. 2.6}$$

$$dDecTr_i = RcDecTr_i \times dTr_i \quad \text{Eq. 2.7}$$

where $age_{Tr_i} =$ age of tracer $i$ (d); $dT_{Tr_i} =$ concentration of conservative tracer $i$ (g m$^{-3}$);

$cTr_i =$ concentration of decayable tracer $i$ (g m$^{-3}$); $RcDecTr_i =$ first order decay rate constant (1 d$^{-1}$); $dDecTr_i =$ flux for decayable tracer $i$ (g m$^{-3}$ d$^{-1}$)

2.3. Results

2.3.1. Calibration

Sediment bed roughness, eddy viscosity coefficient, and wind drag coefficient were varied systematically to calibrate the model. Model water depth at Mike Sub-3 was consistently higher than field measurements (0.15 – 0.30 m, with slightly better agreement before March). This could indicate an issue with the elevation dataset. Comparison of the LiDAR data used and GPS-RTK elevation measurements showed errors in elevation up 20 cm, which could explain some of the water depth differences at these subtidal sites (data not shown). Mike Int-1 and Mike Int-2 exhibited the opposite trend, with the model under-predicting water depth at these stations (0.05 – 0.20 m, with better agreement after March). The uncertainty in sediment bed elevation
highlighted the need for LiDAR data during low tide and low vegetation cover and the need for ground-truthing with a high-resolution GPS – RTK unit.

The final model was chosen to minimize these errors in water depth and temperature at these four platforms. The best results were produced with a piece-wise wind drag function, where coefficient A and B are set at 0.0012 for wind speeds less than 5.6 m s\(^{-1}\) and coefficient C is set at 0.0026 for wind speeds of 14 m s\(^{-1}\). Eddy viscosity coefficients had a relatively minimal effect, but the optimal value was found to be 1.5. Final Manning n values were chosen as 0.01 for channels, 0.02, 0.04, and 0.08, for January-February, March-April, and May-June, respectively, in the subtidal and intertidal floodplains, and 0.20 for all months in the supratidal floodplain.

Normalized root means squared error (NRMSE) for water depth at Mike Int-1, Mike Int-2, Mike Sub-2, and Mike Sub-3 were 14.5%, 19.9%, 12.7%, and 12.7%, respectively for the entire study period. NRMSE values for temperature at Mike Int-1, Mike Int-2, Mike Sub-2, and Mike Sub-3 are 18.6%, 10.5%, 14.5%, and 19.6%, respectively. NRMSE values generally improved during later months as flood stage of river increases (Appendix A. Fig. A.3).

2.3.2. Water Depth, Temperature and Age

Significant differences in water depth, temperature, water storage, and water age occurred across the HGM zones of WLD, responding to seasonality and river stage. In 2015, the river pulse began in mid-March, around the same time that water temperatures increased, and vegetation emerged (Fig. 2.5a). During lower river stage, exchange occurred via low-elevation island edges, predominantly on the distal end of islands, traveled through the island interior, moving from subtidal to intertidal and then to supratidal zones. With lower river stage, water depths were shallower (Fig. 2.5b), especially in supratidal and intertidal floodplains, which led to
larger water temperature fluctuations and warmer daily maximums (Fig. 2.5c). Thus, in winter, supratidal and intertidal HGM zones experienced temperatures greater than 17 °C, which can influence microbial activity. Subtidal floodplains rarely exceeded 12 °C until mid-March.

During low river stage, water spent more time within the island interior, specifically in the supratidal and intertidal floodplains where average water age was 5 days and 2 days, respectively (Fig. 2.5d). However, the water volume within the supratidal and intertidal floodplains was < 4% of the total water within the delta (Fig. 2.5e). Subtidal floodplains maintained a consistent water age of approximately 1 day and contained 1 - 11% of the total water within the delta. Thus, 86 - 98% of the water volume was stored in the primary channels.

From spring to summer, water depths, storage, and temperatures increased across all HGM zones as river stage rose. Over-bank flow delivered water more evenly across the delta producing relatively uniform temperatures and water age between the channels and floodplains. Water age across all HGM zones reduced to <1 day. With an increase in river stage more water flowed into the floodplains; however, the water volume stored in floodplains never exceeded 10% of all water within the delta.
Figure 2.5 (a) Discharge in the Wax Lake Outlet through the simulation period (January – June 2015), (b) average water depth, (c) average water temperature, (d) average water age across each hydrogeomorphic zone, and (e) relative distribution of water volume within the domain.
2.3.3. Exchange and Connectivity

Daily maximums of net channel-floodplain exchange were as high as 1000 m$^3$s$^{-1}$, with occasional peaks of 2000 m$^3$s$^{-1}$ in April (Fig. 2.6a). As expected, net channel-floodplain exchange fluctuated around zero for the edge and volumetric methods and water neither accumulated nor dissipated in the floodplains. Large hourly fluctuations were attributed to tides as they disappear in the no-tides simulation.

Figure 2.6. Net exchange between channels and floodplains (m$^3$/s) within the entire delta using edge method (blue) and fractional exchange (red). Exchange values are shown in two-hour increments. Positive indicates net flow into the floodplains at that point in the simulation and negative indicates net flow out of the floodplains.

To assess hydrodynamic patterns from the delta apex to the delta front, we used the edge method to determine net channel-floodplain exchange within each radial section. In general, both flow into and out of floodplains increased towards the delta front, where natural levees are lower in elevation. While we saw minimal seasonal effects on net channel-floodplain exchange (Fig. 2.5a), flow in and flow out separately indicated a strong river influence, with magnitudes of both flow in and out increasing significantly in March, when the river discharge increased. Net flow
into floodplains occurred in the 4th and 5th radial sections, suggesting water entered the floodplains at these points, traveled through the floodplains, and exited further downstream where net outflow occurs (the 9th and 10th radial sections; Appendix A Fig. A.4).

Net channel-floodplain exchange within each radial section for each season highlight these patterns from the delta apex to the delta front emerged (Fig. 2.7a, solid black line). Near the delta apex, minimal net exchange between channels and floodplains occurred. Net exchange increased 4 km from the delta apex, where both island edge elevations decreased and total edge length increased, thus facilitating more exchange. Maximum flow into floodplains from channels occurred at 8 km with a net flow of 561, 955, 791 m$^3$s$^{-1}$ into the floodplains in winter, spring, and summer, respectively.

Using the edge method, connectivity is the sum of flow into the floodplains relative to sum of flow into the domain via channels (Fig. 2.2). The edge method provided flow into the floodplains and channel flow into the domain is considered at the both the upstream and downstream boundaries. Average connectivity for the entire delta throughout the simulation period was 39.9%. Theoretically, this indicated that 39.9% of the water will interact with the floodplains somewhere within the delta before it is exported out to the delta front. However, due to tides, water may enter and exit multiple times before eventually leaving the domain, therefore complicating the true meaning of connectivity. A time lag between channel flow at the delta apex and the delta front may have produced error in these calculations, but without decomposing the tidal signal, we could not account for this. Connectivity increased from the delta apex to the delta front, with maximum connectivity around 8 km from the delta apex. Within this peak area connectivity, an average of 21%, 19%, and 17% of channel flow moved into floodplains in winter, spring, and summer, respectively (Fig. 2.8b).
Figure 2.7. (a) Net channel-floodplain exchange from the delta apex (0 m) to the delta front (10 km). Positive exchange indicates net transfer of water from channels to floodplains. Negative exchange indicates net transfer of water from floodplains to channels. (b) Connectivity, average percent of channel flow that enters the floodplains. Comparison of vegetation and tide treatments with the original model (solid black line).

To further visualize spatial patterns of exchange across an individual island, flow in and flow out (m$^3$ s$^{-1}$) values were averaged for each edge cell across the entire simulation period (Appendix A Fig. A.5). As expected, lower elevation edges had greater exchange across floodplain edges and are important access points for delivery of water, sediment, and nutrients to the floodplains. In contrast, there was very limited exchange along higher elevation edges of the island. The edges adjacent to primary channels had net flow into the floodplains, while edges facing the delta front, had net flow out.
2.3.4. Effect of Vegetation and Tides

To evaluate the impact of removing vegetation and tides, we compared the net channel-floodplain exchange and connectivity (Fig. 2.9) across the entire delta. Removing tides greatly reduced the large hourly fluctuations in net channel-floodplain exchange that was observed in the original model (Fig. 2.7a). However, removing tides did not change average connectivity across the delta (39.0%). Removing vegetation had a negligible effect on net channel-floodplain exchange, but increased average connectivity to 60.2%. Together the effect of removing tides and vegetation decreased hourly net channel-floodplain exchange, but increased connectivity to 55.2%.

From the delta apex to the delta front, average net channel-floodplain exchange was not changed by removing tides or vegetation (Fig. 2.8a). However, there was a slight increase in exchange (more flow into floodplains) between 6-8 km when vegetation was removed, with a comparable decrease in exchange (more flow out of floodplains) between 9-10 km. However, the effect of these treatment was more apparent in connectivity results, with the most significant differences occurring between 6 and 9 km during spring and summer months. Connectivity peaked at 8 km in all seasons for all treatments. Removing tides reduced connectivity at 8 km by 3% in spring and 6% in summer. Removing vegetation increased connectivity at 8 km by 8% and in spring and 20% summer. Removing tides and vegetation slightly increased connectivity by 4% in spring and 7% in summer.

In addition to the impact on exchange and connectivity, these physical factors also impacted the water volume stored and water age within the floodplains (Fig. 2.8a-b). Removing tides decreased the volume of water stored in the floodplains by 20% in spring and 30% in summer. Removing vegetation decreased the volume of water stored in floodplains by only 5%.
in spring and 10% in summer. Removing tides and vegetation decreased the volume of water stored in floodplains by 26% in spring and 40% in summer. These effects were predominantly observed in the middle sections of the delta, between 6-9 km downstream.

Figure 2.8. Changes in (a) water age and (b) volume of water stored with in floodplains (supratidal, intertidal, and subtidal) in each season, from upstream to downstream.

There was a decrease in water age both from the delta apex to the delta front and from winter to spring (Fig. 2.8). This was expected since water age decreases in subtidal floodplains as river stage rises and subtidal floodplains dominate the downstream portion of the delta. Removing vegetation from floodplains had a negligible effect on water age, while removing tides
decreased water age near the delta apex during the winter and spring, but increased water age in the summer. Removing both vegetation and tides further increased water age during the summer. However, average water age for all floodplains throughout the simulation showed little response to treatment, decreasing from 0.80 days in the original model to 0.79 without vegetation, 0.76 without tides, and 0.76 without vegetation or tides (Fig. 2.9).

![Figure 2.9](image)

Figure 2.9. Changes in (a) connectivity, (b) water age, and (c) water temperature within deltaic floodplains under different tide and vegetation treatments.

2.4. Discussion

Our results support the hypothesis that the three HGM zones of WLD exhibit distinct hydrodynamic patterns in response to river stage and marine forcings. While the tidal range in coastal Louisiana is small (~30 cm), our simulations suggest that tides play a significant role in
connectivity and residence time of coastal deltaic floodplains. From our original hypotheses, we found that flow into deltaic floodplains increased with river stage, but the net channel-floodplain exchange was mostly controlled by tides.

Near the delta apex, natural levees present barriers to flow during low river stage. Therefore, significant exchange occurs on the distal portions of the delta, where floodplain elevations are lower. Unlike tidal marshes which have relatively uniform platforms and steep slopes from marsh to channel, the floodplains at the delta front have a shallow slope from the delta front to the delta apex. When river stage is low, tides maintain hourly exchange along the delta front, a signal that remains even when the river floods. Comparison between the models with and without tides emphasized the role of tides in increasing channel-floodplain exchange. At the interface of river and tide forcings, WLD exhibits bidirectional current flow and like other tidal freshwater systems are dynamic physical and ecological systems with high spatial and temporal variation (Barendregt and Swarth, 2013).

Vegetation establishment and succession during delta evolution creates strong feedbacks between the landscape and hydrologic forcings. Presence of vegetation produces preferential flow along unvegetated areas, therefore controlling the exchange of water between channels and floodplains (Kadlec, 1990; Temmerman et al., 2005a). When vegetation was removed from our model (decreased bed roughness), flow across floodplain edges increased, but water age and water stored within the floodplains was unaffected. This indicates that vegetation reduced the rate of exchange between channels and floodplains, but had less impact on the flow conditions within the floodplains. These results are similar to those found by Nardin et al. (2018), where seagrass and salt marsh vegetation reduced exchange in shallow coastal bay.
However, the use of homogenous vegetation roughness parameters in our model simplified the flow conditions through vegetated areas. In areas with < 40% vegetation cover, such as younger, more distal areas of WLD, Wright et al. (2018) found heterogeneous representation of vegetation allowed for the development of preferential flow paths between channels and floodplains. However, with uniform vegetation parameters, as in our model, these flow paths did not develop, and thus the model may underestimate connectivity in highly vegetated areas. Similarly, vegetation height and density impacted flow velocities through wetlands, which determines sediment trapping (Nardin and Edmonds, 2014), channel erosion (Temmerman et al., 2007), and exposure time (Hiatt et al., 2018). In addition to patchiness, vegetation submergence and stem flexibility can have significant impacts on vegetation resistance to flow, but cannot be accounted for with bed roughness (Khalid and Duan, 2015; Temmerman et al., 2005a). Vegetation type varies significantly at WLD, with large communities of submerged aquatic vegetation in the lower intertidal and subtidal HGM zones (Carle et al., 2015). Incorporation of field-based vegetation parameters could greatly improve our model and better capture the distinct vegetation types across WLD.

Our model estimated a spatiotemporally averaged connectivity of 39.9%, which is within the 24-54% range measured in the field by Hiatt and Passalacqua (2015) between June 16-19, using methods equivalent to the fractional exchange method used in our simulations. In order to make a comparison between our simulation and their field measurements, which do not overlap in time, we chose a portion of our simulation (March 3-5, 2015) that had similar WLO discharge conditions. Hiatt and Passalacqua (2015) completed multiple ADCP transects along two downstream reaches of Gadwall and Main Pass (Fig. 2.10a) during falling and rising tide. During their field measurements the discharge in WLO ranged between 3050-3400 m³s⁻¹.
We averaged discharge measurements from Hiatt and Passalacqua (2015) at each of their transects in Gadwall and Main Pass and compared them with discharge from our simulation at the same transects (Fig. 2.10b). Their measurements and our simulations of discharge were similar in magnitude and both show decreasing channel discharge from the upstream transects to the downstream transects along the island, which indicated increasing connectivity as more water exits the channels and enters the floodplains. We calculated connectivity using the fractional exchange method and found small discrepancies between our simulation and averages of their field measurements, especially between transects L-M and I-J (Fig. 2.10c). We expect these differences derived from resolution in our model, errors in the DEM, and timing of their measurements throughout the tidal cycle.

This comparison suggested that our simulations sufficiently represented the exchange of water between channels and floodplains of WLD. However, it also highlighted potential bias in connectivity measurements depending on location within the delta and time of year. As we discussed above (Fig. 2.5), flow from primary channels to floodplains increased from the delta apex to the delta front due to a reduction in levee elevation. Hiatt and Passalacqua (2015) collected their measurements between 7-10 km from the delta apex, where our simulations show the greatest connectivity between channels and floodplains (Fig 2.8b). With our model, we were able to predict hourly and seasonal changes in exchange and connectivity across WLD, which is important in capturing fluctuations in connectivity over space and time. This is particularly significant given the strong seasonality of river stage and vegetation dynamics. Our model simulations also highlighted the difficulty in defining connectivity as connectivity results had a high sensitivity to the threshold for distinguishing floodplains and channels.
Hiatt and Passalacqua (2015) also measured residence time of Mike Island during February 2014 using a tracer dye experiment. They estimated water remains within the island for at least 3.8 days, which is similar to the average water age we found for Mike Island in February 2015 of our simulation (3.6 days). These results further support our model and the potential for its use in characterizing floodplain hydrology of WLD.

2.5. Conclusion

In summary, water residence time and water temperature vary across the HGM zones of WLD. Higher elevation zones have longer residence times and warmer water temperatures compared to lower elevation zones. However, this trend is specific to season and differences in hydrology among the three HGM zones disappear when river stage increases. Flow of water from the primary channels to the floodplains increases from the delta apex to the delta front and
as river stage rises during seasonal flooding. However, connectivity decreases as river stage rises. Finally, vegetation increases flow into floodplains and increases connectivity, but has little effect on residence time in our simulations, while tides decrease flow into floodplains and connectivity and slightly decrease residence time.

Flow patterns within coastal deltaic floodplains are complex, with interactions of distinct HGM zones with coastal forcings of river, tides and wind complicating model capacity to simulate water residence time and water temperature. Simulating connectivity and residence time are especially important when considering the role of deltaic floodplains in biogeochemical processes such as nitrogen retention and removal. While longer residence times and warmer temperatures in the intertidal and supratidal floodplains create ideal conditions for biogeochemical activity, the relatively small volume of water in these zones could reduce the impact of these HGM zones on biogeochemical cycling of the entire delta. The exchange of water between channels and floodplains is surprisingly large, compared to observations in the field. However, a significant portion of this exchange occurs at the distal ends of each island, where low elevations and very little vegetation present conditions for small changes in water level to create flow into and out of floodplains. Whether most water interacts with the floodplains is still unclear, but will determine the relative contribution of WLD to total nitrate removal (Twilley et al. 2019). Future research will incorporate nitrate removal processes to address nitrate removal potential of coastal deltaic floodplains.
Chapter 3. Spatial Trends in Soil and Vegetation Properties of a Newly Emergent Coastal Deltaic Floodplain

3.1. Introduction

At the initial stages of the coastal delta cycle, water and sediment exit the mouth of a river into an open bay, sediment deposits and the formation of a subaqueous delta begins. Initial deposits in the bay often originates from the river channel bed, producing a sandy subaqueous delta (Roberts et al., 1980). Sediment accumulates as jet-plume deposits, which occur adjacent to and on top of each other to form jet-plume complexes, several of which make up a delta lobe. Eventually the delta becomes subaerial, especially after large flood events, forming a bay head delta, which builds vertically and seaward (Wellner et al., 2005). If conditions for delta growth persist, total area increases and transformation to a shelf delta occurs (Paola et al., 2011). Hydrodynamic forcings including river, wind, tide, and wave energy, as well as sediment load, size, and cohesion, contribute to the development and shape of coastal deltas (Edmonds and Slingerland, 2010; Galloway et al., 2008). River floods and storms contribute to delta development by delivering sediment to floodplains, whereas winter cold fronts resuspend sediment and transport it offshore (Bevington et al., 2017; Turner et al., 2006).

In response to mouth bar formation and channel bifurcation, jet-plume deposits develop into discrete deltaic islands (Edmonds and Slingerland, 2007). These islands, which receive water and sediment from the river and ocean, define the coastal deltaic floodplains (Twilley et al., 2019). Overbank flow produces exchange between channels and floodplains, delivering sediment and nutrients within wetlands (Kadlec, 1990). However, sediment deposition along island edges leads to the formation of natural levees, which reduce overbank flow. Interdistributary bays define the shallow interior and distal portions of deltaic islands, maintaining hydraulic connectivity between channels and floodplains, especially along the delta.
front. Flooding inundation frequency and duration of wetland soils are controlled by sediment elevation as well as hydrologic forcings. Elevation gradients within coastal deltaic floodplains produce distinct hydrogeomorphic zones (HGM), defined by elevation with respect to the local tidal datum (Bevington and Twilley, 2018). Mean low water, mean sea level, and mean high water separate these zones into subtidal, low intertidal, high intertidal, and supratidal.

Through delta evolution and increasing elevations, vegetation establishes and self-organizes across hydrogeomorphic zones (Bevington and Twilley, 2018; Carle et al., 2015). The establishment of vegetation creates strong biological feedbacks as vegetation traps sediment and stimulates organic matter accumulation (Cahoon et al., 2011; Ma et al., 2018; Nyman et al., 1993). Vegetation along floodplain edges trap sediment, especially when intermediate vegetation heights and densities occur, but can also reduce channel-floodplain exchange (Nardin and Edmonds, 2014). Vegetation within the interior of the floodplains reduces flow velocities, increases water residence times, and induces further sediment trapping and nutrient uptake (Kadlec, 1990; Nardin et al., 2016).

Sediment organic matter accumulation rate is the balance of belowground production, organic matter deposition, and decomposition, the latter of which is controlled by environmental conditions including oxygen availability in the soil, limited in inundated wetland systems (Kosters et al., 1987), and salinity (Capone and Kiene, 1988; Ibáñez et al., 2010). Even without mineral sediment supply to island interiors, infilling occurs through organic matter accumulation, producing higher organic soils at higher elevations (Bevington and Twilley, 2018; Lorenzo-Trueba et al., 2012). Sediment bed elevations plateau as sediment delivery is reduced and as soils become aerobic, which balances decomposition and belowground productivity (Cahoon et al., 2011). Therefore, island edge cross-sectional morphology, such as levee width and slope, change
as a function of island age and distance from the island apex (Bevington and Twilley, 2018). Over time, vegetation community distribution and diversity change as species respond to changes in flooding frequency and duration (Johnson et al., 1985; Sasser et al., 1986; Shaffer et al., 1972). Changes in vegetation type and species are good indicators of biological response to the physical environments (Carle et al., 2015; Johnson et al., 1985; Shaffer et al., 1972).

Laboratory experiments highlight the complex interaction between vegetation and water flow as a function of stem structure, flexibility, and submergence (Nepf, 2012, 1999; Nepf and Vivoni, 2000). Even low-density vegetation can significantly increase turbulence, compared with bed-drag and bed-shear, suggesting the importance of vegetation in wetland hydrodynamics (Nepf, 1999). The effect of physical stem attributes can be difficult to simplify, but there are increasing efforts to incorporate them in numerical modeling. More recent improvements in 2D hydrodynamic modeling work to incorporate the influence of vegetation, however mostly through the incorporation of cylindrical rods or increased bed roughness (Al-asadi and Khalid, 2016; Baptist et al., 2007; Klopstra et al., 1997; Nardin and Edmonds, 2014; Straatsma and Baptist, 2008). Several studies have investigated the role of vegetation in decreasing bed shear stress and increasing sedimentation (Nardin et al., 2016), decreasing wave energy and marsh edge erosion (Carr et al., 2018), overall landscape-scale geomorphology (Mariotti and Fagherazzi, 2010; Nardin and Edmonds, 2014), and increasing turbulent kinetic energy (Larsen, 2019; Yang et al., 2016). Specific equations incorporate vegetation height relative to water levels, which determines emergence or submergence, and a frontal area index, referred to as stem number, which is the product of stem density and stem diameter. Application of 3D simulations by Temmerman et al. (2005), where stem structure varied vertically, show even vertical variation in stem structure can be a significant factor flow routing and sediment trapping in tidal marshes.
We explore the role of elevation and age on soil and vegetation properties across a young, prograding coastal delta. We expect strong feedbacks between delta geomorphology and ecology, especially within the initial stages of delta evolution. Sediment bed elevation (classified into HGM zones) controls flooding frequency and duration as a function of connectivity to delta channels as well as delivery of sediment and nutrients. Marsh age (classified as distance from island apex) controls relative contribution of biological and physical processes in the sediment. We target gradients of age and elevation to characterize soil and vegetation properties. We hypothesize that as deltaic islands develop, sediment organic matter content will increase, producing gradients of organic matter content, bulk density, and particle size along an age and elevation gradient. Similarly, we hypothesize that vegetation resistance and aboveground stem biomass increase along an age and elevation gradient. In addition to using these surveys for delta characterization, we develop predictive maps of sediment organic matter content and vegetation structure and explore the sensitivity of hydrodynamic models to these parameters.

3.2. Methods

3.2.1. Study site

Our study site, Wax Lake Delta (WLD), a newly emergent, actively prograding delta in southeastern Louisiana (Fig. 3.1). WLD formed after the dredging of the Wax Lake Outlet by the US Army Corps of Engineers to reduce flooding in Morgan City, Louisiana in 1944. After a large flood in 1973, WLD became subaerial and is estimated to be growing at 1-2 km² hr⁻¹ (Kim et al., 2009) Within WLD, a well-developed island, Mike Island, is the focus of our field surveys. Mike Island is chosen for the defined age chronosequence and the presence of distinct hydrogeomorphic zones. On-site observations as well as previous field measurements show clear changes vegetation, soil properties, surface water flow, and water quality across the island.
Figure 3.1. (a) Louisiana coast within the GOM, (b) Wax Lake Delta classified into hydrogeomorphic zones, identifying (c) Mike Island with soil survey (brown circles) and vegetation survey (black squares) points. (d) Aerial photograph of Mike Island highlighting vegetation changes among hydrogeomorphic zones.

From the apex to the distal tips of delta islands, we observe trends of decreasing elevation, vegetation, and time since emergence. For the past several years, we have been exploring these gradients as a chronosequence, both in river connectivity and soil age (Bevington and Twilley, 2018; Twilley et al., 2019). We classify the island into four HGM zones (subtidal, low intertidal, high intertidal, and supratidal) and four age chronosequence zones. HGM zones are defined by mean high water, mean sea level, and mean low water using USGS Lidar DEM from 2012 (Figure 3.1). Flooding inundation frequency and duration of wetland soils are controlled by sediment elevation as well as hydrologic forcings. In the case of coastal deltaic floodplains such as WLD, river stage is the dominant hydrologic forcing, but tides, winds, and
storm events play an important role as well. Even small changes in sediment bed elevation produce large changes in soil chemistry and vegetation composition.

The four chronosequence zones (originally described in Bevington and Twilley, 2018) are generally defined by the distance from the island apex and characterized by distinct changes in cross-sectional morphology. Chronosequence I, near the island apex, is comprised of mostly older, supratidal HGM, and generally higher soil organic matter sediments. Sediment infilling from belowground productivity produces relatively flat bed elevations. Chronosequence II transitions to more intertidal (high and low) zones with supratidal levees along the outer edges of the island. Around chronosequence III and IV low intertidal and subtidal HGM zones dominate with generally younger and lower organic matter content sediments. Soils are sandy and vegetation is sparser, with a transition from emergent to submerged species from chronosequence III to IV.

### 3.2.2. Soil Survey

In spring 2015, we conducted a soil survey across Mike Island to characterize the soil properties and nutrient storage of WLD. Instead of four chronosequence zones, we used seven age transects for the soil survey, to capture transitions within the four age chronosequence zones. Transects are not evenly spaced along the island, but were instead chosen to capture expected changes in soil properties (Fig. 3.1). The sampling design focused on the island interior, thus there was an uneven distribution of sites among HGM zones. To increase the number of supratidal cores, we included additional cores collected along the island levees when testing the HGM hypotheses.

A small (2.37-cm diameter) piston corer was used to collect 15 cm sediment cores, which were sectioned in 1-cm increments in the top 5 cm and 5-cm increments in the bottom 10 cm.
Core sections were dried at 60° C to determine bulk density. Particle size was determined on a subsample from each dried section using a Beckman-Coulter laser particle size analyzer (Model LS 13 320). The remaining sample was ground and used to determine organic matter content using loss on ignition (Heiri et al., 2001). Sediment porosity, the amount of pore space within the sediment, was calculated as the percent of core volume that was not occupied by organic or mineral mass.

### 3.2.3. Vegetation Survey

In spring and summer 2017, we conducted a vegetation stem survey across Mike Island to characterize vegetation structure and biomass of WLD. The vegetation survey used the original four chronosequence zones and did not incorporate the distal ends of the island where emergent vegetation is not present. Three sampling events captured initial plant emergence after winter (March/April), mid-summer (June), and peak biomass (September/October). Sample points were randomly selected within each HGM zone across specific chronosequence zones. The vegetation survey was designed to represent both hydrogeomorphic zone and chronosequence. However, due to the geomorphology of the delta, certain zones were not available in each chronosequence. For example, subtidal and low intertidal zones do not occur in chronosequence I and supratidal zone do not occur in chronosequence IV.

We targeted two dominant species, *Nelumbo lutea* in the supratidal and high intertidal HGMs, and *Colocasia esculenta* in the subtidal and low intertidal HGMs (Fig. 3.2). Non-dominant species comprised a small percent of stem density at most sites and were excluded from analyses. Quadrats with 4m² and 1m² dimensions were used for the *N. lutea* and *C. esculenta* sites, respectively, due to the low density of stems in the *N. lutea* sites. Stems were clipped and brought back to the lab, where stems of dominant species were counted, diameters
were measured with calipers at 20-cm increments along each stem, and stem heights were recorded. Stems were dried to constant weight at 60° C to determine aboveground biomass.

![Image of vegetation]  

Figure 3.2. Dominant species used in this study (a) *Colocasia esculenta* and (b) *Nelumbo lutea*

We measured vegetation structure to determine how stem characteristics potentially impact floodplain hydrodynamics of WLD. Specifically, we measured stem density (m, # stems m⁻²), stem diameter (d, every 20cm from the base of the tip of each stem), stem number (n, the product of diameter and density), stem volume (v, the sum of volume of each 20 cm section, calculated as a cylinder), and basal area (a, the sum of stem area per sediment area). While *N. lutea* showed relatively consistent, cylindrical structures with little change in diameter from the base to the tip, *C. esculenta* had a tapered shape. Therefore, basal diameters decrease significantly from average diameters, especially as the plant grows, thus we consider both basal and average stem number throughout this study.

### 3.2.4. Prediction Maps

Denitrification is an important pathway for nitrate removal in coastal wetlands and has been shown to increase with sediment organic matter content (Arango et al., 2007; Henry and
Therefore, estimating landscape-scale denitrification is limited, in part, by a lack data on spatial variability of sediment properties. Soil core data from this study were combined with soil core data available from previous studies at WLD (Aarons, 2019; Shields et al., 2017) to build simple and multilinear regression models of sediment organic matter content using independent variables. Variables include sediment bed elevation and wetland age, which is often approximated using distance from the delta apex (Henry and Twilley, 2014). Previous studies show that sediment organic matter content increases with elevation (Bevington and Twilley, 2018; Shields et al., 2019). The advantage of these variables is the ability to extract them across the entire delta, using remote sensing or simple geometry. We normalized distance from the delta apex and distance from individual island apexes (Fig. 3.3a-b). Sediment bed elevations were derived from USGS Lidar DEM from 2012 (NAVD88, Fig. 3.3c). In addition, we included horizontal distance from channel (Fig. 3.3d), as the sediment deposition has been shown to decrease away from the marsh edge (Reed et al., 1999; Temmerman et al., 2003). We tested each variable and combination to find the best model for predicting sediment organic matter content of WLD.
In Chapter 2, we introduced a Delft3D hydrodynamic model of WLD, which was used to study flow patterns, connectivity, and water residence time within the floodplains of the delta. For the original model, a simple Manning’s $n$ was used to simulate vegetation and bed roughness. Manning’s $n$ was applied to supratidal, intertidal, and subtidal hydrogeomorphic zones and those values were adjusted during calibration for three seasons of the simulation (winter, spring, and summer). However, interest in the impact of vegetation on connectivity and residence time led to improving the representation of vegetation within the model. Using vegetation parameters from this survey, we built simple maps of vegetation structure using significant trends found in our survey and applying average values based on HGM zone, species,
and chronosequence zone. These vegetation structural maps were input into our original Delft3D hydrodynamics model to determine the sensitivity of water depth and connectivity to stem structure parameters.

We chose to use the Baptist (2007) equations for roughness, which incorporate stem height, stem number, drag coefficient, and water depth into calculation of a Chezy roughness value. Vegetation is still incorporated as a bed roughness coefficient, but with an added flow resistance term \((-\frac{\lambda}{2}u^2)\) in the momentum equation.

For submerged vegetation, Chezy roughness coefficient is modeled as

\[
C = C_b + \frac{\sqrt{g}}{\kappa} \ln \left( \frac{h}{h_v} \right) \sqrt{1 + \frac{C_p n h_v C_b^2}{2g}}
\]

Eq. 3.1

and the flow resistance is modeled

\[
\lambda = C_d n \frac{h_v}{h} \frac{c_b^2}{c^2}
\]

Eq. 3.2

where \(n\) is stem number \((n = mD)\), \(m\) is stem density, \(D\) is stem diameter, \(C_d\) is drag coefficient \((1)\), \(C_b\) is the Chezy bed roughness \((45 \text{ m}^{1/3} \text{ s}^{-1})\), \(h\) stem height, \(h_v\) is water depth, \(g\) is gravitational acceleration, and \(K\) is the Von Karman Constant \((0.4)\). When stem height is greater than water depth and vegetation becomes emergent, this equation 1 becomes equation 3 and equation 2 becomes equation 4.

\[
C = C_b
\]

Eq. 3.3

\[
\lambda = C_d n
\]

Eq. 3.4

One thing to note regarding the Baptist equations is that when the vegetation is emergent, the roughness parameter becomes bed roughness, the same parameter included as part of the original model. The effect of emergent vegetation is instead reflected as a flow resistance in the momentum equation. For more details, reference Deltares (2016).
3.3. Results

3.3.1. Soil Characteristics

Overall, sediment bulk density and particle size increase, while sediment organic matter content and porosity decrease along the chronosequence within the top 15cm (Fig. 3.3). Bulk density increased with depth as the organic fraction decreases. Particle size remained relatively constant with depth with slight increases in some transects. Soil porosity decreased with depth due to compaction at greater soil depths.

We sampled seven sections of the soil column, but limited our statistical analysis to the surface for testing hypotheses of age and elevation gradients. It is within this top layer that most microbially mediated biogeochemical processes occur and where sediment is first deposited. To decide how many soil sections to pool, we compared ANOVA results for the top 0-1 cm, 0-2 cm, 2-4 cm, and 0-4 cm. All sections that included the top 1 cm (0-1 cm, 0-2 cm, and 0-4 cm) show statistically significant differences for all soil characteristics along the age chronosequence. The 2-4 cm sections only shown differences in bulk density and particle size. Only the top 1 cm showed significant differences across HGM zones and only for organic matter content. Ultimately, we pooled 0-2 cm and 2-4 cm separately and present them independently to examine changes in trends at different depths.
Figure 3.4. Soil depth profiles averaged along the chronosequence for (a) bulk density, (b) organic matter content, (c) particle size, and (d) porosity.

Within the 0-2 cm sections, there was a significant increase in bulk density along the chronosequence ($p < 0.001$), a simultaneous decrease in organic matter content ($p < 0.010$) and porosity ($p = 0.002$), and an increase in particle size, or decrease in phi ($p = 0.007$) (Fig. 3.4). Following a Tukey HSD post-hoc comparison, most differences in bulk density, porosity, and particle size occurred between transect 1 and transects 6 and 7. Differences in organic matter content occurred between transect 1 and all other transects (Fig. 3.4). These trends are attributable to a transition from physical to biological processes controlling soil development.

There was also a decrease in elevation at these sampling points and an increase in connectivity to the primary distributary channels of the delta, which produces sandier sediments in the distal end of the island. However, sediment organic matter density (g cm$^{-3}$) was highest in the middle of the delta where the balance between total mass and organic fraction is maximized. With such low bulk density and high porosity in first chronosequence of the island, there was very little total mass, either organic or inorganic.
Within the 2-4 cm segments, there is a significant increase in bulk density ($p = 0.006$), and particle size ($p = 0.010$) along the chronosequence, but a significant decrease in organic matter content ($p < 0.001$) and porosity ($p = 0.011$) along the chronosequence. (Fig. 3.5). As with the 0-2 cm section, Tukey HSD post-hoc comparisons show that differences mostly occur between transect 1 and transects 6 and 7. In the case of porosity, there are no pair-wise differences.
Figure 3.6. Chronosequence changes within 2-4 cm of (a) bulk density, (b) organic matter content, (c) particle size, and (d) porosity

Sample distribution among HGM was uneven, with more samples at lower elevations. The only statistically significant difference between HGM zones was in the 0-2 cm section for organic matter content ($p = 0.038$), with the supratidal zone having the highest organic fraction (Fig. 3.6 and Fig. 3.7). While not significant, bulk density decreases with increasing elevation, while porosity increases in both 0-2 and 2-4 cm sections. This lack of significant differences across HGM zones indicated the role of mineral sediment delivery and connectivity within the delta channels. However, in supratidal zones, the influence of the river lessens and biological
feedbacks between vegetation and soil led to greater accumulation of organic matter within the sediment.

Figure 3.7. Hydrogeomorphic zone differences within the top 0-2cm of (a) bulk density, (b) organic matter content, (c) porosity, d) particle size
3.3.2. Vegetation Characteristics

Despite targeting peak biomass in September/October, we did not measure large increases in *C. esculenta* biomass between seasons (Appendix Fig. B.2). At older sites (chronosequence zone I), *C. esculenta* biomass decreased from March to June and from March to September/October. *N. lutea* biomass increased at all sites, with much greater productivity between the June and September/October sampling events. Within each chronosequence, *C. esculenta* biomass production was greater than *N. lutea*. Net productivity was calculated as the
difference in biomass (g m\(^{-2}\)) between March and June or September, for \(C. \text{esculenta}\) and \(N. \text{lutea}\) sites, respectively (Fig. 3.9).

![Figure 3.9. Net primary productivity, calculated as the change in biomass over time, for each hydrogeomorphic and chronosequence zone](image)

There were significant differences in all vegetation structure variables between the dominant species (\(p < 0.001\)). \(C. \text{esculenta}\) grew in denser stands at higher elevations, whereas \(N. \text{lutea}\) stands were sparser, with thinner stems and a single leaf. For each species, month had a significant effect on stem height (\(p < 0.001\)), stem density (\(p = 0.012\) for \(C. \text{esculenta}\) and \(p < 0.001\) for \(N. \text{lutea}\)), and stem volume (\(p < 0.001\)). Basal area, average stem number, basal stem number, and biomass were statistically different among months for \(N. \text{lutea}\) (\(p < 0.001\)), but not for \(C. \text{esculenta}\) (\(p = 0.717\), \(p = 0.288\), \(p = 0.288\), and \(p = 0.428\)). Therefore, when considering hydrogeomorphic zones and chronosequence, we analyzed each species and month separately.
By isolating each species, we reduced the number of HGM and chronosequence zones for each species to two and three, respectively.

In general, most differences between HGM and chronosequence zones occur in June and September/October months. In March, HGM only had a significant effect on stem heights of *C. esculenta*, while chronosequence had a significant effect on stem height (p<0.0001), stem density (p=0.027), basal area (p=0.0065), stem volume (p=0.0059), and biomass (p<0.0001) of *C. esculenta*. For *N. lutea* the only significant effect of chronosequence was on stem height (p<0.0001) and biomass (p=0.0066) (Appendix B. Fig. B. 1-4). In June, HGM had no significant effect on any vegetation variables of *N. lutea*. However, all variables except stem height were significantly different across HGM for *C. esculenta* (p = 0.0480, p = 0.0414, p = 0.005, p = 0.0099, p = 0.0043, p = 0.019 for stem density, average stem number, basal area, stem volume, basal stem number, and biomass). Stem heights were significantly different across chronosequence for both species (p = 0.015 and p = 0.007 for *C. esculenta* and *N. lutea*, respectively). In September/October, there were no significant differences between HGM or chronosequence zones for *C. esculenta* for any variables. For *N. lutea*, HGM zones were significantly different for average stem number, basal area, stem volume, basal stem number, and biomass (p = 0.011, p = 0.013, p = 0.019, p = 0.025, p = 0.011, respectively).
Figure 3.10. Differences in basal stem number between species, age chronosequence, and season for (a) supratidal and high intertidal hydrogeomorphic zones, and (b) low intertidal and subtidal hydrogeomorphic zones.

3.3.3. Prediction Maps

3.3.3.1. Sediment Organic Matter Content

We found a strong correlation between sediment organic matter content and sediment elevation, distance from delta apex, and distance from island apex, each normalized using minimum and maximum values measured across WLD ($r^2 = 0.67 \ p < 0.001$, $r^2 = 0.68 \ p < 0.001$, $r^2 = 0.32 \ p < 0.001$, respectively). There was no correlation with sediment organic matter and normalized distance from channel ($r^2 = 0.01 \ p = 0.862$). However, elevation and normalized distance were also correlated with each other, with one variable explaining 62% of the variance of the other (Fig. 3.11).
Figure 3.11. Correlated bed elevation and normalized distance from island apex, showing trends in soil organic matter content.

The spatial heterogeneity of sample locations within this study was limited, comprising mostly island interiors. To incorporate other islands and increase spatial extent, we compiled data from other WLD sources (Fig. 3.12; Aarons, 2019; Shields et al., 2016; Bevington and Twilley, 2018; Twilley et al. unpublished data). Due to the deeper collection depths of these cores (10-15cm), we used 10 cm of the cores collected in this study for this analysis. Dates of collection vary between 2010 and 2017.
Figure 3.12. Additional soil cores used in the sediment organic matter models to increase spatial coverage of data.

With increased spatial heterogeneity provided by the increased sample size, we saw a decrease in the variance explained by each of the model variables, with less than 45% explained by any variable. Normalized distance to delta apex and sediment bed elevation had the best correlations to organic matter content (Fig. 3.13 a-b). Cores from Bevington and Twilley (2018) were collected in transects from the channel edge and capture changes along the natural levees. However, the organic matter trends in those cores diverged considerably from those of this study, possibly suggesting different factors controlling soil properties of interior and levee environments.
Figure 3.13. Linear regressions between sediment organic matter content (%) (A) normalized distance from delta apex and (B) bed elevation.

We chose four models that incorporate 1) only natural log transformed normalized distance from delta apex ($r^2 = 0.424$, $p < 0.001$), 2) only untransformed normalized sediment bed elevation ($r^2 = 0.204$, $p < 0.001$), 3) all four natural log transformed variables ($r^2 = 0.444$, $p$ <
0.001), and 4) all four untransformed variables \(r^2 = 0.412, p < 0.001\) (Fig. 3.14a-d). Each variable in the fourth model is significant, but normalized distance from delta apex had the largest effect. While the root mean squared error is similar for all models, Model 3 has the largest \(r^2\) and lowest root mean squared error (3.11), but only two variables are significant (Table 3.1).

Table 3.1. Model parameters and statistics for predicting sediment organic matter

<table>
<thead>
<tr>
<th>Model</th>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
<th>Model 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model</td>
<td>(\ln(OM) = 0.222 - 1.388 \ln(A))</td>
<td>(\ln(OM) = 0.718 + 2.322C)</td>
<td>(\ln(OM) = 0.268 - 1.601 \ln(A) + 0.132 \ln(B) + 0.023 \ln(D) - 0.65 \ln(C))</td>
<td>(\ln(OM) = 2.285 - 3.221A + 0.554B + 0.659D + 0.876C)</td>
</tr>
<tr>
<td>(r^2)</td>
<td>0.424</td>
<td>0.204</td>
<td>0.441</td>
<td>0.412</td>
</tr>
<tr>
<td>p-value</td>
<td>&lt;0.001</td>
<td>&lt;0.001</td>
<td>&lt;0.001 ((A &lt;0.001, B = 0.018, C = 0.483, D = 0.940))</td>
<td>&lt;0.001 ((A &lt;0.001, B = 0.034, C = 0.033, D = 0.045))</td>
</tr>
<tr>
<td>RMSE</td>
<td>3.13</td>
<td>3.62</td>
<td>3.11</td>
<td>3.19</td>
</tr>
</tbody>
</table>

\(A = \) Normalized Distance from Delta Apex

\(B = \) Normalized Distance from Island Apex

\(C = \) Normalized Sediment Bed Elevation m NAVD88

\(D = \) Normalized Horizontal Distance from Channel

These methods have significant limitations, particularly due to the small number of soil cores used in developing models and arising from assumptions regarding similarity of island geomorphology and soil development. Ecologically, both variables are important for determining organic matter content, in addition to biological feedbacks that are not currently available at larger spatial scales. However, these maps can be used to estimate spatial heterogeneity of biogeochemical processes, most importantly denitrification, across WLD.
3.3.3.2. Vegetation Structure

We are interested in determining the impact of small changes in the structure of dominant vegetation on the hydrodynamics of WLD. While we found that different species, and thus different hydrogeomorphic zones, have statistically significant differences in stem height, stem density, and stem number, the importance of these differences is unknown. The most significant differences in vegetation structure occur between species, but we also measure differences (not always significant) by HGM and chronosequence. To test the sensitivity of our model to these parameters, we had four scenarios for predicting stem height and basal stem number (Table 3.1). Each scenario applied stem height and basal stem number from this survey. The first scenario...
used average values by species in June, 78.2 and 105.6 cm stem height, and 2.09 and 0.02 m $m^2$ for basal stem number of *C. esculenta* and *N. lutea*, respectively. The second scenario used average values by HGM zones in June of 86.9, 72.4, 101.8, 109.3 cm stem height and 2.515, 1.807, 0.019, 0.028 m $m^2$ for basal stem number, for supratidal, high intertidal, low intertidal, and subtidal HGM zones, respectively. The third scenario used average values for chronosequence zones for each species in June. For *C. esculenta*, stem heights are 88.6, 74.6, 64.8 cm and basal stem numbers are 1.97, 2.31, 1.90 m $m^2$ for chronosequence I, II, and III respectively. For *N. lutea*, stem heights are 85.9, 114.6, 116.2 cm and basal stem numbers are 0.04, 0.02, 0.02 m $m^2$ for chronosequence II, III, and IV, respectively.

Table 3.2. Vegetation structure parameters for model sensitivity analysis

<table>
<thead>
<tr>
<th>Scenario 1</th>
<th>Stem Height (cm)</th>
<th>Basal Stem Number (m $m^2$)</th>
<th>Scenario 2</th>
<th>Stem Height (cm)</th>
<th>Basal Stem Number (m $m^2$)</th>
<th>Scenario 3</th>
<th>Stem Height (cm)</th>
<th>Basal Stem Number (m $m^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>N. lutea</em></td>
<td>105.6</td>
<td>0.023</td>
<td>Supratidal</td>
<td>86.9</td>
<td>2.515</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>C. esculenta</em></td>
<td>78.2</td>
<td>2.09</td>
<td>High Intertidal</td>
<td>72.4</td>
<td>1.807</td>
<td><em>N. lutea</em> - Chronosequence I / II</td>
<td>85.9</td>
<td>0.035</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Low Intertidal</td>
<td>101.8</td>
<td>1.019</td>
<td><em>N. lutea</em> - Chronosequence III</td>
<td>114.6</td>
<td>0.016</td>
</tr>
<tr>
<td>Subtidal</td>
<td>109.3</td>
<td>0.027</td>
<td></td>
<td></td>
<td></td>
<td><em>C. esculenta</em> - Chronosequence I</td>
<td>88.6</td>
<td>1.965</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><em>C. esculenta</em> - Chronosequence II</td>
<td>74.6</td>
<td>2.314</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><em>C. esculenta</em> - Chronosequence III / IV</td>
<td>64.8</td>
<td>1.893</td>
</tr>
</tbody>
</table>

The fourth scenario applied stem heights and basal stem numbers based on empirical relationships between biomass and these vegetation structure variables. Thomas et al. 2019
extrapolated biomass values for WLD based on Sentinel 2 data and field measurements from Oct 2016. From our vegetation survey, we developed empirical relationships between vegetation structure and biomass from our Sept/Oct dataset. We found a significant relationship between stem heights and biomass for *C. esculenta* (*p* < 0.0001), but not for *N. lutea*. Instead, the average stem height for Sept/Oct for *N. lutea* was applied (79.1 cm). We found significant relationships between basal stem number and biomass for both species (*p* < 0.0001 and *p* = 0.0009 for *N. lutea* and *C. esculenta*, respectively). Unfortunately, there are significant areas of the delta that do not have biomass data and forested areas that have biomass values much greater than we measured in our survey. When data was missing, mostly in the subtidal HGM zones and chronosequence 4, we applied average stem heights and basal stem numbers by species. We also capped stem heights and basal stem numbers at our largest measured values to prevent unreasonable values within the forested areas where Thomas et al. 2019 predicted very large biomass values. Even though this biomass map was available only for October 2016, we applied the resulting stem height and basal stem number maps in our June simulation for this sensitivity analysis. For each of these three scenarios we produced predictive maps of stem height and basal stem number (Fig. 3.16 and Fig.3.17). Using the Delft3D trachytotes parameter, we defined two, four, eight, and twenty-two trachytotes for the four scenarios.
Figure 3.15. Prediction maps of stem height based on four classifications: a) by species, b) by hydrogeomorphic zones, c) by chronosequence and species, and d) as a function of biomass.

Figure 3.16. Prediction maps of basal stem number based on four classifications: a) by species, b) by hydrogeomorphic zone, c) by chronosequence and species, and d) as a function of biomass.
To test the sensitivity of WLD hydrodynamics to vegetation structure, we compared water levels at a station within the interior of Mike Island and net channel-floodplain exchange within a three-week simulation predicted by the baseline model of Chapter 2 and four modified models with different ways of parameterizing heterogeneous vegetation roughness. For the simulation period of June 1 – 21, 2015, we used vegetation parameters from the June/July field survey.

Water depth at an interior subtidal HGM station decreased up to 20 cm in each modified model (Fig. 3.18a). The chronosequence simulation, which assigns vegetation parameters based on distance from island apex and species, had the most similar water depths to the original model, differing by less than 10 cm. The other three models were very similar to each other. All vegetation models except the model based on biomass demonstrated greater flow from channels to floodplains between 4-8 km and greater flow from floodplains to channels between 9-11 km (Fig. 3.18b). The introduction of the Baptist representation in the Delft3D model was substantially different than the original model, which only used sediment bed roughness.
Figure 3.17. Water Depths at interior, subtidal site on Mike Island, comparison of four modified vegetation models with the original model from chapter 2.

3.4. Discussion

Throughout the growth phase of the delta cycle, sediment delivery allowed for soil accretion and progradation (Roberts, 1997). At these initial stages, soil development was dominated by sediment delivery from primary channels, producing sandy, low organic matter soils. However, as elevations increased and vegetation established, belowground productivity increased soil organic matter content (Fisk, 1952; Lorenzo-Trueba et al., 2012). Our results evidenced this transition with decreasing bulk density and increasing organic matter content along the chronosequence of delta islands. Continued accretion allows productive coastal
marshes to compensate for rising sea levels, allowing for the possibility of continued progradation under climate change scenarios.

Field surveys of soil and vegetation highlight spatial trends on delta development related to transition from physical to biological processes and the feedback between the two. Trends from the island apex to the island front most clearly highlight soil development from mineral to increasingly organic, though the organic fraction remained low when compared with older Louisiana wetlands (Wang et al., 2017). This transition was associated with both age and connectivity to the deltaic channels, which increased from transect 1 to transect 7. Even within deeper sediments (15-30 cm) Aarons (2019) found patterns of increasing organic matter content and decreasing bulk density, as well as increasing nitrogen to phosphorus ratios and total carbon.

Aside from species differences, trends in vegetation along the chronosequence are less clear. With greater water depths, taller *N. lutea* maintains emergence. Feedbacks between trapping efficiency and mean elevation of each species suggests an ecological feedback between plant location and resulting impacts on island development. While we focused on these two dominant species, forests and submerged aquatic vegetation are both important vegetation communities that should be considered in the HGM zones.

The sensitivity analysis of our hydrodynamic model to vegetation structure parameters suggests that including vegetation parameters is important, but that small changes in vegetation height and stem number do not produce large effects on floodplain connectivity. Therefore, it is possible to apply these parameters to functional groups across a larger landscape. Vegetation parameters can be extrapolated through remote sensing, incorporating publicly available radar and hyperspectral data (Byrd et al., 2018; Straatsma and Middelkoop, 2007) into hydrodynamic models (Straatsma and Baptist, 2008). Normalized difference vegetation index and soil-adjusted
vegetation index can also improve extrapolation of vegetation structure to the landscape scale. With enough field data, vegetation type, height, and density can be simplified for an entire ecosystem, even herbaceous marshes. Further exploration into developing these tools could help build robust hydrodynamic models and improve our understanding of biological and physical feedbacks in these important coastal systems.

3.5. Conclusions

Overall, results from this research have improved our understanding of surface processes of an active coastal deltaic floodplain. For both soil and vegetation, chronosequence is a better predictor of spatial and seasonal trends, which may indicate the role of channel-floodplain connectivity during early stages of delta development. While elevation has a significant effect on many soil and vegetation properties, the differences mostly occur between the highest elevations (supratidal HGM zone) and all other lower elevations. Therefore, we conclude that the relative influence of physical (river, tide, and wave energy) and biological (vegetation composition and productivity) factors on soil and vegetation properties changes over time, with some time threshold that initiates a transition to biological processes. Soil and vegetation interactions are key for predicting the development of WLD and can also be useful in predicting the outcome of proposed river sediment diversions. Further development of relationships for soil and vegetation properties with remotely sensed data could significantly change our ability to incorporate physical and biological processes into numerical models. While predictive maps are simple, they provide an additional tool for improving hydrodynamic and water quality models.
Chapter 4. Quantifying the Fate of Nitrate in a Coastal Deltaic Floodplain Using an Isotope Enrichment Field Experiment

4.1. Introduction

Since the 1960s, nitrogen loading to the Mississippi River watershed has increased more than two-fold (Goolsby and Battaglin, 2001; Rabalais et al., 2002; Turner and Rabalais, 1991), largely as a result of agricultural and urban runoff (Burkart and James, 1999). An increasing percent of this nitrogen load is nitrate-nitrogen, which is bio-available and readily used within an ecosystem. Increases in nitrogen loading are also attributed to changes in precipitation, which affects the amount of applied fertilizer that drains into the rivers, and discharge of the tributaries of the Mississippi River. Draining more than 40% of the continental United States, and up to 950,000 Mg of nitrate-N, the Mississippi River delivers large nitrate loads to the Gulf of Mexico, which nourish phytoplankton communities (Murphy et al., 2013). Large phytoplankton blooms sink to the benthic zone reducing bottom-water dissolved oxygen concentrations, as the decomposition of such a large population consumes significant amounts of oxygen in the water. Eutrophication and eventually hypoxia occur just off the Louisiana Bight, with summer areal extents increasing each year (Rabalais et al., 2002).

Opportunities for nitrate removal exist throughout the Mississippi River watershed and continue out along the Louisiana coast (Mitsch et al., 2005a, 2001). Plant and microbial uptake, denitrification, and other microbially mediated processes remove or retain nitrate temporarily and permanently. While a fraction of the retention occurs within the river itself, a significant amount occurs within the wetlands and estuaries of the river system. Alluvial floodplains are nitrate-limited and can thus have a significant impact on nitrate removal (Richardson et al., 2004). In the Chesapeake Bay, Noe and Hupp (2009) found that increased floodplain size and flood inundation time could produce nitrate retention of up to 24% of riverine load. However, in
the Atchafalaya River, a distributary of the Mississippi River, BryantMason et al. (2013) found that only 7% of river nitrate was removed. One of the largest limitations on nitrate removal in alluvial floodplains is hydrologic connectivity. Increased channelization of the Mississippi River has limited the exchange between river and floodplain and thus the delivery of nitrate to floodplain wetlands.

Within the depositional zone of the Mississippi River watershed, both the Mississippi River Delta and the Louisiana estuaries have the potential for nitrate removal as well. For example, while BryantMason et al. (2013) quantified nitrate removal within the Atchafalaya River, Lane et al. (2002) found much greater removal (41 - 47%) when considering the distal end of the system, including the Atchafalaya Delta and Bay. Due to their higher soil organic matter content, shallow depths, warm temperatures, and longer residence times, coastal marshes can play a large role in nitrate removal. However, while significant rates of denitrification occur in active deltas (Henry and Twilley, 2014), the larger size of coastal estuaries allow for equal or greater contribution of these estuaries to total nitrate removal (Perez et al., 2011).

The fate of nitrate discharged from a major river depends on physical, biological, and chemical factors. The most explored pathway for nitrate removal is denitrification, which removes nitrate by releasing it to the atmosphere as dinitrogen gas, eliminating it as a source of nitrogen in eutrophication. However, other biogeochemical pathways have been shown to play a significant role in wetlands including dissimilatory nitrate reduction to ammonium (Burgin and Hamilton, 2007), which converts nitrate to ammonium, another bioavailable form of nitrogen.

With improved methods of measuring stable isotopes of nitrogen, tracing the heavy nitrogen isotope, $^{15}$N, has become a very important technique for studying nitrogen within ecosystems. Using labeled $^{15}$N allows scientists to directly measure different pathways of the
nitrogen cycle through core incubations (An and Gardner, 2002; Groffman et al., 2006; Henry and Twilley, 2014; Hou et al., 2012; Steingruber and Friedrich, 2001), mesocosms (Cornelisen and Thomas, 2002; White and Howes, 1994; Wozniak et al., 2008), and even landscape-scale experiments (Gribsholt et al., 2007). Each experimental scale produces a range of control and realism, contributing to a greater database on nitrogen cycling. Beyond just producing rates in different ecosystem, labeled $^{15}$N experiments have also been used to study the effect of flow hydrodynamics on ammonium and nitrate uptake in sea grass beds (Cornelisen and Thomas, 2004; Thomas et al., 2017), long-term retention of nitrogen in tidal fresh and salt marsh (Drake et al., 2008; White and Howes, 1994), nitrogen uptake by microphytobenthos (Eyre et al., 2016; Veuger et al., 2007), and nitrate removal in streams (Mulholland et al., 2009; O’Brien et al., 2012) and urban wetlands (Harrison et al., 2012).

We studied the transformation of nitrate in a freshwater wetland to quantify the removal capacity of newly emergent coastal deltaic floodplains. Mesocosm field experiments were designed with two stages: 1) a pulse of $^{15}$NO$_3$ tracer to the water column and hourly monitoring 2) a chase of $^{15}$N tracer through vegetation and soil compartments over several months. The pulse-chase experiments were designed to test the effect of soil organic matter content on short-term and long-term nitrogen cycling and fate of surface water nitrate. We hypothesize that sites with higher organic matter content will have greater rates of denitrification. The pulse-chase experimental design allows for an initial pulse of stable isotope tracer, $^{15}$N in this case, followed by monitoring of the pathways through which the tracer moves. The use of stable isotopes isolates the added tracer from ambient nitrate in the system and verifies the tracer as the source of measured $^{15}$N within other components throughout the experiment. Control reference samples account for ambient concentrations of $^{15}$N within each component.
4.2. Methods

4.2.1. Study Site

Our study site is within a well-developed island of Wax Lake Delta (WLD), a young, actively prograding delta in southeastern Louisiana. WLD formed as a result of a USACE flood risk reduction project, which dredged the Wax Lake Outlet. Sediment deposited at the mouth of the outlet and after a particularly large flood in 1973, WLD became subaerial. WLD is predicted to be growing at a rate of 1-2 km$^2$ yr$^{-1}$ (Kim et al., 2009). Several islands make up WLD, with Mike Island located in the middle, oriented north-south and having distinct elevation ranges and vegetation communities. We chose Mike Island for this nitrate experiment because of the well-developed geomorphology, patterns of organic matter and soil elevation, and significant preliminary data on water quality, vegetation communities, and soil biogeochemistry (Fig. 4.1).

We identified three sites for mesocosm experiments with low, medium, and high soil organic matter content (Lower-OM, Med-OM, and Higher-OM, respectively Table 4.1.) based on initial loss on ignition measurements within the top five centimeters of soil. The Lower-OM and Higher-OM sites, located in low intertidal and subtidal zones, respectively, are dominated by *Nelumbo lutea*. Med-OM, located in the supratidal zone, is dominated by *Colocasia esculenta*. 
Figure 4.1. Maps showing a) coastal Louisiana (insert shows the Gulf of Mexico), b) our study site, Wax Lake Delta, classified into hydrogeomorphic zones based on USGS National Elevation Dataset Lidar Elevations, and c) the locations of mesocosm experiments and associated photos. D) Mesocosm design with triplicate bottomless aluminum tanks, 1.16 m in diameter and 1.5 m tall, pushed into 0.1-0.2m into the soil.

4.2.2. Pilot Study

In 2017 and 2018, we conducted two pilot studies to test the application of our $^{15}$NO$_3$ tracer as well as our conservation tracers, NaCl and propane. The first experiments were done in two field flumes built as part of another project (Larsen et al., in prep). However, due to high cost of $^{15}$NO$_3$ tracer, we excluded flow, effectively producing a chamber (10 m x 1 m). The second experiments were done in a portable flume (4 m x 1 m), where we tried to create 5cm/s flow of water. Experiments were run in subtidal, low intertidal, or high intertidal vegetation zones with submerged aquatic vegetation, *N. lutea*, or *C. esculenta*, respectively. While the pilot
study demonstrated our ability to detect N\textsubscript{2} isotopes over a one-day incubation, it also highlighted issues with leakage through the flume seams and the porewater as well as the incredible difficulty in producing flow for the duration of the experiment. After the pilot study, we redesigned our mesocosms to be seamless circular, bottomless tanks. This new design prevented any leakage through the mesocosm walls and allowed us to insert the mesocosms further into the marsh sediment. While we were not able to accomplish 5 cm s\textsuperscript{-1} current, we did add a pump to help circulate water within the mesocosms. For a full description of mesocosm design and methods testing see Appendix A.

4.2.3. Field Mesocosm Design

Triplicate mesocosm enclosures were built for field incubation experiments to test the role of sediment organic matter content on the fate of nitrate in coastal deltaic floodplains. These enclosures were designed to minimize edge effect and incorporate sufficient area to represent the plot-scale ecology. Enclosures were constructed using 0.32 cm aluminum sheets welded into a cylinder (1.16 m diameter) and inserted > 10 cm into the soil. A small propeller pump circulated the water in each enclosure to help mix the water column, but flows were very small and not evenly distributed in the water column. A generator supplied power to the pumps throughout the multi-day experiments and marine batteries supplied power to the instruments and datalogger.

Within each mesocosm at each site, we carried out two short-term and one long-term experiment. During the short-term experiments \textsuperscript{15}NO\textsubscript{3} tracer was added and monitored for several days. The mesocosms were left in place between each pulse experiment and for one to two weeks following the second pulse to allow for maximum inoculation of the tracer into the sediment. The long-term experiment began after the mesocosms were removed and continued for two to six months, until October 2019. The long-term experiments aim to chase the \textsuperscript{15}N pulse
that entered the soil and plants during the short-term experiments. During these months, we sampled storage compartments including aboveground vegetation, belowground vegetation, and sediment.

On July 12, 2019 Hurricane Barry passed over WLD, producing strong winds and 1.2-1.8 m of storm surge (Fig. 4.2). The effect of Hurricane Barry was seen across the delta (Fig. 4.3), killing most of the emergent vegetation, which returned in some areas, but not at our study sites (Fig. 4.4). Unfortunately, without vegetation after July, we were unable to sample vegetation $^{15}$N throughout the entire growing season. Instead, we only have aboveground vegetation data until June or July 2019.

![Figure 4.2. From Cangialosi et al. 2019 (National Hurricane Center) Estimated maximum storm surge inundation levels (ft. above ground) along the northern Gulf coast during Hurricane Barry. Estimates are based on NOS tide station data above Mean Higher High Water, USGS streamflow gauges, and a SLOSH hindcast. Image courtesy of the NHC Storm Surge Unit.](image)
Figure 4.3. A) Vegetation browning after Hurricane Barry on Mike Island. B) Damage to trees along the levees of Mike Island.
4.2.4. Short-Term “Pulse” Experiments

At the beginning of each short-term pulse, we added a $\text{K}^{15}\text{NO}_3$ isotope tracer (enrichment between 30-80 μM $^{15}\text{NO}_3$) and NaCl conservative salt tracer (30 mg L$^{-1}$). After adding the tracer, we manually mixed the water column of each mesocosm for 10-15 minutes, while simultaneously bubbling propane as a soluble gas conservative tracer. Surface water samples were collected before the injection to set a baseline (T0) and to calculate initial isotopic enrichment. NaCl concentrations are used as a conservative tracer to account for any leakage.
through groundwater exchange and propane concentrations are used to estimate water-atmosphere exchange of N$_2$.

Slightly different sampling schedules occurred for each experiment. For the first site, Med-OM (April 2019), the 1$^{\text{st}}$ pulse had a target enrichment of 30 μM $^{15}$NO$_3$ and the 2$^{\text{nd}}$ pulse had a target enrichment of 70 μM $^{15}$NO$_3$. Both pulses were sampled only on the day of injection and occurred 3 days apart. For the second two sites, Lower-OM and Higher-OM (May and June 2019), both the 1$^{\text{st}}$ and 2$^{\text{nd}}$ pulses had target enrichments of 45 μM $^{15}$NO$_3$. For these latter sites, the 1$^{\text{st}}$ pulse was sampled one hour after injection and throughout the following day. The 2$^{\text{nd}}$ pulse occurred three to four days later, with an injection in the morning and consecutive sampling for the next four to five hours.

During each sampling event water level, temperature, conductivity, and dissolved oxygen were measured using a meter stick, salinometer (YSI 30SCT), and dissolved oxygen probe (HACH HQ30D), respectively. During the Lower-OM and Higher-OM experiment, two to three mesocosms were equipped with pressure transducers (Campbell CS 451), conductivity and temperature probes (Campbell CS 547A), and surface water nitrate sensors (Seabird Scientific Submersible Ultraviolet Nitrate Analyzer SUNA V2) to make continuous measurements of water depth, temperature, conductivity, and nitrate/nitrite. Unfortunately, we were limited in the number of instruments and experienced technical difficulties during periods of each experiment. However, these measurements helped us to identify environmental changes overnight and during days when we were not sampling.

We collected 250 mL of surface water (just below the surface), which was stored on ice, filtered within 24 hours, and frozen for inorganic nutrient analysis (NO$_2$, NO$_3$, NH$_4$, PO$_4$, Si). Whatman GFF (0.45 μm) filters were used and saved for chlorophyll a analysis. In addition, we
used a syringe to slowly collect surface water for N\textsubscript{2} gas analysis with minimal aeration to limit
the amount of contamination from air. N\textsubscript{2} gas samples were collected in 12 mL exetainers,
allowing two volumes to overflow, preserved with concentrated ZnCl\textsubscript{2}, and stored under water at
4°C. We also collected 100 mL of surface water for total nitrogen, which were frozen unfiltered.
Three 40 mL glass vials were filled with 30 mL of surface water with airtight septa caps for
measuring propane and stored at 4°C. Finally, at select sampling events, porewater samples were
collected using a porewater sipper at approximately 3 cm soil depth and filtered for inorganic
nutrient analysis (NO\textsubscript{2}, NO\textsubscript{3}, NH\textsubscript{4}, PO\textsubscript{4}, Si).

NaCl and propane samples are used to correct any loss of \textsuperscript{15}N through air-water gas
exchange, groundwater leakage, precipitation, or evaporation. Relative conductivity is calculated
using conductivity immediately following tracer addition and is used to correct all nutrient
measurements for loss of tracer. Initial conductivity for all sites is <0.2 ppt or 200 μS cm\textsuperscript{-1}.
Propane samples were collected at each sampling event and used to determine an air-water
exchange rate coefficient to correct for loss of \textsuperscript{15}N through diffusion. Air-water exchange
coefficients (k\textsubscript{p}) for propane were calculated as the slope of the linear regression of the natural
log of propane. Peak area of propane was determined with an SRI 861C Gas Chromatograph
with a Flame Ionization Detector. To determine an N\textsubscript{2} air-water exchange coefficient, k\textsubscript{n2} was
calculated to correct N\textsubscript{2} flux rates according to

\[
k_{n2} = k_p \left( \frac{D_{n2}}{D_p} \right)^x
\]

where D\textsubscript{n2} and D\textsubscript{p} are the diffusion coefficients for N\textsubscript{2} and propane, respectively and x is the
Schmidt number dependence, which for smooth waters is estimated to be -2/3 (Jähne et al.,
1987). Diffusion coefficients were interpolated from values from Wise and Houghton (1966) for
a water temperature of 27 °C and found to be 3.5 and 2.4 x10\textsuperscript{-5} cm\textsuperscript{2} sec\textsuperscript{-1} for N\textsubscript{2} and propane,
respectively. From this calculation, $K_{n2}$ is 0.778 times $k_p$. Unfortunately, due to a broken cold room, several propane samples from the first two experiments do not have detectable propane. Therefore, we apply propane air-water exchange rates from Higher-OM to all our sites.

Filtered surface and porewater samples for nutrient analysis were run on an OI Analytical Auto-Analyzer for NO$_2$, NO$_2$+NO$_3$, NH$_4$, and PO$_4$. Rates were calculated as the change in concentration over time, using concentrations that were corrected for conservative tracer loss. Unfiltered samples were oxidized using a persulfate digestion to convert particulate and dissolved organic nitrogen to NO$_3$. These samples were run on the auto-analyzer for NO$_2$+NO$_3$ to determine total nitrogen and organic nitrogen was calculated as total nitrogen minus NO$_x$ and NH$_4$. Using discrete water samples, we calculated NO$_x$ fluxes for the 1$^{st}$ pulse (night and day) and 2$^{nd}$ pulse (day only) at the Lower-OM and Higher-OM sites. Night fluxes are calculated as the change in concentration between the end of the pulse day (~5:00pm) and the start of the following day (~9:00am), corrected with conservative tracer values. Day rates are calculated using sampling events between 9:00am and 4:00pm. Due to timing of field days and increased leakage over multiple hours (full tidal cycle) at Med-OM, we only calculated day rates for the 2nd pulse.

Isotope analysis was run on NH$_4$ using the OX-MIMS method (Yin et al. 2014), which reduces NH$_4$ to N$_2$ using a sodium hypobromide solution. While some labile organic nitrogen is also converted in this method, we expect this is small relative to the production of $^{15}$NH$_4$. Dissimilatory nitrate reduction to ammonium is calculated based on the production of $^{15}$NH$_4$ and atom% $^{15}$N enrichment of the surface water nitrate as

$$DNRA = p^{15}NH_4 \times \frac{(^{15}NO_3+^{14}NO_3)}{^{15}NO_3}$$

Eq. 4.2.
We were unable to develop an appropriate method for determining $^{15}\text{NO}_3$ concentrations and therefore must assume constant enrichment of surface water nitrate. Any fractionation through nitrate uptake processes is not considered.

Dissolved N$_2$ samples were analyzed using a Membrane Inlet Mass Spectrometer to determine concentrations of $^{29}\text{N}_2$ and $^{30}\text{N}_2$. Calculations for denitrification are based on the isotope pairing technique (Nielsen, 1992). Denitrification of $^{15}\text{NO}_3$ ($D_{15}$) is calculated as the production of $^{29}\text{N}_2$ ($r_{29}$) plus two times the production of $^{30}\text{N}_2$ ($r_{30}$), corrected for NaCl and propane loss.

$$D_{15} = r_{29} + 2r_{30} \quad \text{Eq. 4.3.}$$

Denitrification of $^{14}\text{NO}_3$ ($D_{14}$) and total denitrification ($D_{\text{total}}$) are calculated as

$$D_{14} = \frac{r_{29}}{2r_{30}} + D_{15} \quad \text{Eq. 4.4.}$$

$$D_{\text{total}} = D_{14} + D_{15} \quad \text{Eq. 4.5.}$$

Assumptions of the isotope pairing technique include a linear increase in denitrification with increasing nitrate concentration, no interference of added $^{15}\text{NO}_3$ nitrate to $^{14}\text{NO}_3$ denitrification, homogeneous mixing within the water column and sediment of $^{15}\text{N}$ and $^{14}\text{N}$, and a stable tracer gradient across the sediment water interface. However, these assumptions were not tested directly in our systems due to logistics of performing multiple field experiments. Therefore, we use the isotope pairing technique only to estimate total denitrification and do not distinguish between direct and coupled nitrification-denitrification. Also, the presence of field conditions including vegetation (secretion of O$_2$ around roots) and bioturbation would likely make the assumptions of homogeneity invalid. However, when calculating mass balance (see below), we use only $^{15}\text{N}$ rates for all processes.
4.2.5. Long-Term “Chase” Experiments

The short-term experiments occurred over three to four days, after which mesocosms were left in place for two to four weeks to allow greater incorporation of the tracer into the sediment porewater. At the completion of the short-term experiments, enclosures were removed and the long-term experiments to chase the $^{15}$N isotope pulse began. For each chase sampling event, three cores were taken. The first core, 6-in diameter down to 50cm, was taken over a plant stem to target belowground roots and rhizomes. The aboveground stems were clipped from the belowground portion in the field. The belowground portion was washed over a 2 mm sieve to remove mineral material and sorted into live and degrading organic material. Samples were dried at 60°C until constant weight, ground, and run for carbon and nitrogen analysis on a CHNS-O Elemental Combustion System (Costech 4010) and $^{15}$N/$^{14}$N isotope ratios on a Gas Chromatograph Isotope Ratio Mass Spectrometer (Europa Scientific, now Sercon, 2020).

The second two cores, 3.7 cm diameter down to 10cm, were collected to determine total $^{15}$N and $^{15}$NH$_4$. Both cores were divided into 1-cm sections down to 5 cm. One core was oven dried at 60 °C to constant weight, ground, and used for determining bulk density, organic matter content (Heiri et al., 2001), carbon to nitrogen ratios, and total soil $^{15}$N. The second core was stored frozen for porewater and exactable $^{15}$NH$_4$ analysis, using 8-12 g of wet soil, extracted with 50 mL 2M potassium chloride. From nutrient analysis, no samples showed sufficient extractable NO$_3$ to warrant isotopic analysis – thus $^{15}$N was determined for NH$_4$ only. The OX-MIMS method was used to determine $^{15}$NH$_4$ concentrations from potassium chloride extracts.

We report isotopic enrichment in the delta notation (%o) with respect to the international standard of atmospheric nitrogen (Air):

$$
\delta^{15}N \text{(%o)}_{\text{Air}} = \left[ \frac{R_{\text{Sample}}}{R_{\text{Standard}}} - 1 \right] \times 1000 \quad \text{Eq 4.6.}
$$
In addition to cores, we also clipped two vegetation plots nearby to determine biomass (g m\(^{-2}\)), stem density (# m\(^{-2}\)), and average stem heights (cm). In addition to site characterization, these measurements were also used to extrapolating \(15\)N measured within cores on an areal basis.

**4.2.6. \(15\)N Mass Balance**

Throughout the chase experiments, we inventoried the mass of \(15\)N stored within each compartment for a total mass balance. To account for ambient \(15\)N concentrations, we calculated atom\% \(15\)N excess as the difference between atom\% \(15\)N measured in mesocosm samples and control reference samples. When reference samples were not available, we subtracted atom\% \(15\)N values from T0 samples taken prior to pulse experiments. The mass of \(15\)N produced from our pulse experiments was calculated according as

\[
\text{Mass } 15\text{N} = \text{atom}\% 15\text{N} \times \%\text{N} \times \frac{\text{mass of component}}{\text{core}}
\]

Eq. 4.7.

For each component, we extrapolated mass in cores to the entire mesocosm footprint using best estimates of biomass per area. For aboveground components, we used nearby vegetation plots (g m\(^{-2}\)) to extrapolate \(15\)N mass to the full mesocosm area. This is especially important for the species at WLD, which have relatively sparse densities compared to salt marshes. For belowground roots and rhizomes, we extrapolated to the mesocosm using root:shoot ratios within the cores, multiplied by the biomass from the vegetation plots to approximate the belowground live biomass per m\(^2\). We assumed homogeneous areal distribution of belowground degrading organic material. Therefore, we extrapolated degrading organic material to the mesocosm using the mass per m\(^2\) measured in the cores.

Soil components are calculated similarly, using bulk density (g cm\(^{-3}\)) to determine the T\(^{15}\)N mass within the footprint of the mesocosm at each soil core depth. Final mass balance
values are presented in mg $^{15}$N within each compartment throughout the chase experiments (four – six months).

### 4.2.7. Disturbance Experiments

After Hurricane Barry, we added a core incubation experiment to look at the effect of disturbance on benthic fluxes. In August 2019, we collected sediment cores from each mesocosm at the Lower-OM, Med-OM, and Higher-OM. Our flow-through incubation setup included a continuous flow of 100 μM ambient NO$_3$ site water, filtered from the Wax Lake Outlet. Each core contained 20 cm of sediment, collected to minimize disturbance, with a 10 cm water column. Flow rates were set to produce a 3-hour residence time of overlying water within each core and cores were held in a 21°C water bath. After a pre-incubation period of 9 hours, samples for nutrient and N$_2$ gas analysis were collected every three hours for nine hours under dark conditions. Surface water was filtered for inorganic nutrient analysis; no $^{15}$NH$_4$ was detected using the OX-MIMS method. N$_2$ dissolved gas samples were run on the MIMS using the N$_2$/Ar technique. Following the dark incubation, 14.2 μmoles s$^{-1}$ LED lights (449 - 630 nm) were placed over the chamber and cores were again sampled every three hours for nine hours.

Between the dark and light incubations, one sediment core from each site was sacrificed and used to collected two small (2.37 cm) and one tiny (1 cm) sediment cores. Small cores were sectioned into 1-cm intervals down to 5 cm and 5-cm interval down to 10 cm for bulk density, loss on ignition, total $^{15}$N, and extractable $^{15}$NH$_4$ analysis. Tiny cores were separated into 0-2 and 2-4 cm and frozen for chlorophyll a analysis. A set of control cores were run under the same conditions to correct for any fluxes that occur within the water column or other experimental effects.
4.3. Results

4.3.1. Site Characterization

Before each experiment began, site characterization included initial measurements of soil, vegetation, and water parameters. Bulk density and organic matter content confirmed classification of each site into lower, medium, and higher organic matter with 4, 7 and 10% organic fraction, respectively (Table 1). However, it should be noted that our Higher-OM site is high relative to organic matter content measured across Mike Island, but is lower than many older coastal marshes where the organic fraction comprises up to 60% (Wang et al., 2017). There is little variation in bulk density or organic matter content within the top 5 cm of soil and we pool the 0-1 and 1-2 cm soil portions as well as the 2-3 and 3-4 sections of each core (Fig. 4.5).

![Figure 4.5](image.jpg)

Figure 4.5. Sediment profiles of a) bulk density, and b) soil organic matter content, down to 10cm. Error bars represents standard error based on three soil cores.

In 2019, water levels at WLD were high throughout the spring, delaying the emergence of *N. lutea* and producing low water temperatures until April. Initial daytime water temperatures
at the Lower-OM, Med-OM, and Higher-OM sites were 24, 19, and 28 °C, respectively. Due to differences in sediment bed elevation, as well as temporal changes in tide and river stage, initial water depths varied by 20-70 cm at the three sites. The shallowest water depths occurred at Med-OM (30 cm), followed by Lower-OM (80 cm), and Higher-OM (100 cm). The tides (~30 cm) produced minor variations in water level within each mesocosm due to hydrostatic pressure between the inside and outside of the mesocosm. Data from the available pressure transducers highlight minor fluctuations in water levels within the mesocosms. Morning dissolved oxygen concentrations decreased through the season as water temperatures increased, with the lowest values measured at Higher-OM in June. In fact, Higher-OM experienced anoxic conditions, an important mesocosm effect that may impact the nitrogen pathways measured at this site. These effects are discussed in more detail below.

The Lower-OM and Higher-OM sites were dominated by *N. lutea*, with a small amount of submerged aquatic vegetation. Vegetation measurements (Table 1) are presented for June 2019, before Hurricane Barry hit, to make comparisons between all three sites. In June, Lower-OM had greater biomass, but smaller stem densities (295 g cm⁻³ and 20 stems m⁻²) than Higher-OM (208 g cm⁻³ and 29 stems m⁻²). Stem heights were slightly taller at Higher-OM (122 cm) than Lower-OM (106 cm), likely due to greater water depths. Med-OM was dominated by *C. esculenta*, with some *A. philoxeroides* and, after Hurricane Barry, *Sagittaria latifolia*. In June, Med-OM had the greatest biomass and stem density (443 g m⁻³ and 81.1 stems m⁻²) of all sites, with the shortest stem heights (96 cm). Chlorophyll a concentrations were similar across sites at the initiation of the short-term experiments (3.6, 5.6, and 2.6 mg L⁻¹ at Lower-OM, Med-OM, and Higher, respectively). However, chlorophyll a Lower-OM increased throughout both pulses indicating some growth stimulation from the nitrate additions.
Table 4.1. Site characterization of Lower-OM, Med-OM, and Higher-OM sites based on measurements made during mesocosm deployment unless otherwise indicated. Vegetation data are from June and July 2019.

<table>
<thead>
<tr>
<th>Site</th>
<th>Initial Water Column</th>
<th>Vegetation Plots (June-July, before Hurricane Barry)</th>
<th>Soil Cores</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Depth (cm)</td>
<td>Temp (°C)</td>
<td>Dissolved Oxygen (mg L⁻¹)</td>
</tr>
<tr>
<td>Lower-OM (May)</td>
<td>72</td>
<td>24</td>
<td>6.7</td>
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<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Med-OM (April)</td>
<td>35</td>
<td>19</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Higher-OM (June)</td>
<td>98</td>
<td>28</td>
<td>5.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>
4.3.2. Short-Term “Pulse” Experiments

4.3.2.1. Conservative Tracers

NaCl tracers were injected at approximately 30 mg L$^{-1}$ as a dissolved conservative tracer. In all experiments, specific conductance was monitored at each sampling period. In general, the relative conductivity did not drop below 94% during each short-term experiment (Appendix A, Fig. A.1). At Med-OM, high sediment porosity, due to a dense root structure, increased porewater exchange and inhibited complete isolation of the mesocosms from surrounding water. There was also a rain event after the 1$^{st}$ pulse that diluted the conservative tracer after the 1$^{st}$ pulse. Between the 1$^{st}$ and 2$^{nd}$ pulse, all NaCl tracer was lost (or was highly diluted with rainwater), implying complete loss of $^{15}$N tracer over a full tidal cycle. We only incorporated 2$^{nd}$ pulse data for calculations of flux since there was no loss of NaCl during the five-hour experiment.

During some sampling events, the relative conductivity increased, suggesting either that the water column was not well mixed or there was some addition of salt to increase the conductivity of the water. From the Lower-OM experiments, we noticed that as tides changed and water levels in the mesocosm fluctuated, there was a simultaneous change in conductivity. As water in the mesocosms rose, indicating porewater transport across the sediment-water interface into the water column, conductivity increased. As water in the mesocosms fell, indicating surface water moving to the porewater, conductivity decreased. This could indicate that increases in conductivity were due to the introduction of NaCl, or other salt, from the porewater. NaCl and non-conservative components such as NO$_3$ may be held in the porewater during low tide and reintroduced to the mesocosms as tides rise. While it is impossible to quantify the exact exchange of NaCl between surface and porewater, we speculated from our
results that less than 4% of the conservative tracer was lost across any given experiment. We used these loss estimates to correct all nutrient data below.

4.3.2.2. Inorganic Nutrient and Organic Nitrogen Fluxes

Depletion of dissolved oxygen and surface water nitrate at Higher-OM indicated strong mesocosm effects, predominantly during the 2\textsuperscript{nd} pulse. This effect was less obvious at Lower-OM, where aerobic conditions remain, and nitrate was not depleted throughout the short-term experiments. However, we measured significant changes in nitrate uptake rates during the 2\textsuperscript{nd} pulse and suspected some mesocosm effects at Lower-OM as well. Dissolved oxygen concentrations remained \textgreater 5.4 mg L\textsuperscript{-1} at Med-OM and starting nitrate concentrations (50-62 \textmu M) did not indicate nitrogen limitation in the water column. As we expected 2\textsuperscript{nd} day pulses at Lower-OM and Higher-OM to have strong mesocosm effects, we averaged only 1\textsuperscript{st} pulse day and night to test our hypotheses. However, all fluxes are presented for discussion.

At Lower-OM, nitrate plus nitrite concentrations, measured in one mesocosm equipped with a SUNA, were nearly constant right after the 1\textsuperscript{st} pulse (Fig. 4.6). From the morning following the 1\textsuperscript{st} pulse to just before our 2\textsuperscript{nd} pulse (~48 hours), there is a linear decrease in NO\textsubscript{x} concentration, calculated as a flux of -660 \textmu moles NO\textsubscript{x} m\textsuperscript{-3} hr\textsuperscript{-1}. After the 2\textsuperscript{nd} pulse, the SUNA recorded five hours of NO\textsubscript{x} data, which with a flux of -2110 \textmu moles NO\textsubscript{x} m\textsuperscript{-3} hr\textsuperscript{-1}. 
For the Higher-OM experiments, we deployed SUNAs in all three mesocosms, but instrument error produced large gaps in NO\textsubscript{X} data for two mesocosms during the 1\textsuperscript{st} pulse. Unlike Lower-OM, nitrate plus nitrite concentrations decreased shortly after tracer addition (Fig. 4.7). NO\textsubscript{X} concentrations decreased linearly for the first 24 hours as concentrations decreased from 65 μM to 10 μM NO\textsubscript{X}, with a flux rate of -2490 μmoles NO\textsubscript{X} m\textsuperscript{3} hr\textsuperscript{-1}. After 24 hours, flux signal became smaller than instrument noise, likely due in part to limitations of the SUNA. During the first five hours after the 2\textsuperscript{nd} pulse, the average flux of all mesocosms was -7051 μmoles NO\textsubscript{X} m\textsuperscript{3} hr\textsuperscript{-1}, which was larger than the 1\textsuperscript{st} pulse, a trend we found in the discrete water samples as well.
NO₃ fluxes calculated from discrete water samples indicated nitrate plus nitrite removal occurred even within the first night of the experiment. Available SUNA NO₃ rates compared well with discrete sample rates (Fig. 4.8). Night rates, measured after the 1ˢᵗ pulse, were significantly larger (p = 0.003) at Higher-OM (-2474 μmoles m⁻³ hr⁻¹) than Lower-OM (-505 μmoles m⁻³ hr⁻¹). In one Higher-OM mesocosm, NO₃ was undetectable on the morning after the 1ˢᵗ pulse and is excluded from all 1ˢᵗ pulse calculations. Day NO₃ fluxes at Lower-OM increased from -282 after the 1ˢᵗ pulse to -2405 μmoles m⁻³ hr⁻¹ after the 2ⁿᵈ pulse. Day NO₃ fluxes at Higher-OM increased from -1566 after the 1ˢᵗ pulse to -6058 μmoles m⁻³ hr⁻¹ after the 2ⁿᵈ pulse. Significant overnight uptake produced very low NO₃ concentrations in the Higher-OM mesocosms and nitrogen limitation in the water column may be responsible for the large increase in nitrate fluxes during the 2ⁿᵈ pulse. Also, the dissolved oxygen levels were very low at Higher-OM, allowing for more anaerobic processes, such as denitrification. Med-OM 2ⁿᵈ pulse day NO₃ fluxes were -2371 μmoles m⁻³ hr⁻¹. Including only experiments without strong mesocosm effects (1ˢᵗ pulse day/night average for Lower-OM and Higher-OM and 2ⁿᵈ pulse day for Med-OM) ANOVA results indicated a significant effect of site on NO₃ fluxes (p = 0.015), with Tukey HSD pairwise comparisons indicating significant differences between Lower-OM and the other two sites, but not between Med-OM and Higher-OM.

NH₄ fluxes were much smaller than NO₃ fluxes (Table 2, Appendix A. Fig. A.2). Night and day fluxes of the 1ˢᵗ pulse at Lower-OM were similar (-187 and -170 μmoles m⁻³ hr⁻¹, respectively) but decreased to -32 μmoles m⁻³ hr⁻¹ during the 2ⁿᵈ pulse, likely due to very low NH₄ availability during the 2ⁿᵈ pulse. Night NH₄ fluxes at Higher-OM showed initial consumption of available NH₄ (-38 μmoles m⁻³ hr⁻¹). However, day fluxes showed NH₄ production, rather than consumption with larger fluxes occurring after the 2ⁿᵈ pulse (1199 μmoles
m$^{-3}$ hr$^{-1}$) than after the 1$^{st}$ pulse (703 μmoles m$^{-3}$ hr$^{-1}$). 2$^{nd}$ pulse day NH$_4$ fluxes at Med-OM were not significantly different from 0, suggesting little to no change in NH$_4$ in these experiments. There was a significant effect of site (again excluding 2$^{nd}$ pulse Lower-OM and Higher-OM, and 1$^{st}$ pulse Med-OM) on NH$_4$ fluxes (p < 0.001), with Tukey HSD pairwise comparisons indicating significant differences among all three sites.

![Figure 4.8: Nitrate + Nitrite fluxes, averaged across three mesocosm replicates. Error bars show standard error.](image)

Total nitrogen fluxes were generally similar to dissolved inorganic nitrogen fluxes (NH$_4$ + NOx), suggesting small organic nitrogen fluxes (Table 2). We calculated organic nitrogen fluxes as the difference in total and dissolved inorganic nitrogen fluxes. However, inconsistent organic nitrogen concentrations between replicates prevented calculation of rates. While organic
nitrogen production likely occurred as nitrate uptake by microbial assimilation, we were not able to estimate this process.

PO₄ fluxes at Lower-OM and Med-OM were negative and relatively constant (-29 μmoles m⁻³ hr⁻¹ and -38 μmoles m⁻³ hr⁻¹, respectively). Night PO₄ fluxes at Higher-OM were also similar (-27 μmoles m⁻³ hr⁻¹), but became positive during 1st pulse and 2nd pulse day experiments (50 and 104 μmoles m⁻³ hr⁻¹, respectively). However, high variability between the mesocosms produced large standard errors. There was a significant effect of site (again excluding 2nd pulse Lower-OM and Higher-OM, and 1st pulse Med-OM) on PO₄ fluxes (p = 0.009), with Tukey HSD pairwise comparisons indicating significant differences between Higher-OM and the other two sites, but not between Lower-OM and Med-OM.

Table 4.2. Inorganic nutrient and organic nitrogen fluxes (μmoles m⁻² hr⁻¹) measured across Lower-OM, Med-OM, and Higher-OM triplicate mesocosms. Standard errors are indicated in parentheses.

<table>
<thead>
<tr>
<th>Site</th>
<th>Pulse</th>
<th>Denitrification</th>
<th>NH₄</th>
<th>NOₓ</th>
<th>SUNA</th>
<th>TN</th>
<th>TON</th>
<th>DNRA SW</th>
<th>DNRA PW</th>
<th>PO₄</th>
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<td></td>
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<tr>
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<td>(158)</td>
<td>(204)</td>
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<td>(2)</td>
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<td>1st Pulse</td>
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<td>-1475</td>
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<td>(Day)</td>
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<td>(259)</td>
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<td>(10)</td>
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<td>5965</td>
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<td>(Day)</td>
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<td>(674)</td>
<td>(671)</td>
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<td>(76)</td>
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<tr>
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<td>1199</td>
<td>-6058</td>
<td>-7051</td>
<td>-5103</td>
<td>628</td>
<td>281</td>
<td>(181)</td>
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<td>(1344)</td>
<td>(1389)</td>
<td></td>
<td>(66)</td>
<td>104</td>
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</table>
4.3.2.3. Denitrification

After correcting for air-water gas exchange and Na Cl loss, denitrification rates were determined for each of the short-term experiments at each site (Fig. 4.9). Highest denitrification rates at Lower-OM occurred during the 1\textsuperscript{st} pulse night experiments (832 μmoles N m\(^{-3}\) hr\(^{-1}\)) and then decreased to 658 and 396 μmoles N m\(^{-3}\) hr\(^{-1}\) during the 1\textsuperscript{st} and 2\textsuperscript{nd} pulse day experiments. 2\textsuperscript{nd} pulse day N\(_2\) production rates at Med-OM were 618 μmoles N m\(^{-3}\) hr\(^{-1}\). For Higher-OM, the 1\textsuperscript{st} pulse night N\(_2\) production rate was 1841 μmoles N m\(^{-3}\) hr\(^{-1}\), which decreased to 1389 μmoles N m\(^{-3}\) hr\(^{-1}\) the following day. After the 2\textsuperscript{nd} pulse, N\(_2\) production increased significantly with a day rate of 5257 μmoles N m\(^{-3}\) hr\(^{-1}\). This sharp increase in N\(_2\) production may be related to the anoxic conditions and sudden supply of NO\(_3\) that encouraged denitrifying bacteria. There is a significant effect of site on denitrification (p < 0.001), with Tukey HSD pairwise comparisons indicating significant differences between Higher-OM and the other two sites, but not between Med-OM and Lower-OM.
4.3.2.4. Dissimilatory Nitrate Reduction to Ammonium

\(^{15}\text{NH}_4\) production, DNRA, was measured throughout each short-term experiment, in the surface water and porewater (Fig. 4.10). \(^{15}\text{NH}_4\) fluxes more directly indicated that the source of \(\text{NH}_4\), produced through DNRA was \(\text{NO}_3\). From our ambient \(\text{NH}_4\) fluxes, which were only positive in Higher-OM, surface water \(\text{NH}_4\) production only occurs in Higher-OM. The \(^{15}\text{NH}_4\) concentrations support this finding as we did not detect any \(^{15}\text{NH}_4\) in the surface waters of the Lower-OM and Med-OM experiments. At Higher-OM, 1\text{st} and 2\text{nd} pulse day surface water DNRA rates were 126 and 281 \(\mu\text{moles N m}^{-3}\text{ hr}^{-1}\), which were lower than ambient \(\text{NH}_4\) fluxes. Therefore, another process could be producing \(\text{NH}_4\) in the surface water, likely the decomposition of organic nitrogen.

Figure 4.9. Denitrification rates for each short-term experiment. Error bars show standard error.
$^{15}$NH$_4$ concentrations in the porewater were very low at all sites, likely due to slow diffusion of $^{15}$NO$_3$ through the sediment water interface or low DNRA rates in the porewater. Porewater rates were only calculated for the day samples. The 1$^{st}$ pulse and 2$^{nd}$ pulse Lower-OM porewater DNRA day rates were 429 and 414 μmoles m$^{-3}$ hr$^{-1}$, respectively, which were similar to the 2$^{nd}$ pulse day rates at Med-OM of 516 μmoles m$^{-3}$ hr$^{-1}$. The 1$^{st}$ pulse day rates at Higher-OM were positive, but smaller (317 μmoles m$^{-3}$ hr$^{-1}$). After the 2$^{nd}$ pulse there was a decrease in porewater $^{15}$NH$_4$ (-90 μmoles m$^{-3}$ hr$^{-1}$) indicating negligible DNRA and instead uptake of ammonium within the porewater. With significant uptake of NO$_3$ in the surface water in Higher-OM, the diffusion of $^{15}$NO$_3$ in the porewater may be reduced. There was no significant effect of site on porewater DNRA rates ($p = 0.238$).

Figure 4.10: Dissimilatory nitrate reductin to ammonium rates for each short term experiment. Error bars show standard error.
4.3.2.5. Short-term Mass Balance

Using rates for $^{15}$N, we compared all processes accounted for in the short-term experiments (Fig 4.11). Total nitrate fluxes were larger than the sum of all other processes at Higher-OM and Med-OM. From the NaCl measurements, we didn’t expect significant $^{15}$NO$_3$ loss during the period of flux calculations. Therefore, another process, likely water column assimilation or transfer to the sediment, must be occurring. However, at Lower-OM, total nitrate fluxes were smaller than denitrification and DNRA, when porewater $^{15}$NH$_4$ production was included. Overall, changes in tracer concentrations in Lower-OM were small and the imbalance could be related to sampling error, calculations, or assumptions about $^{15}$N:$^{14}$N ratios.

![Figure 4.11: Comparison of total $^{15}$NO$_x$ fluxes with denitrification and DNRA highlighting a major process that is removing nitrate and nitrite from the mesocosms.](image-url)
4.3.3. Long-Term “Chase” Experiments

As mentioned above, Hurricane Barry hit on July 12, 2019, producing conditions that killed the plants near our study sites. With this destruction, the long-term chase of our $^{15}$N tracer was cut short and led to complications in predicting storage within the soil and vegetation compartments. At the Lower-OM, Med-OM, and Higher-OM sites, Hurricane Barry occurred after our 2-month, 3-month, and 2-week sampling event, respectively. Therefore, our Higher-OM site has the most limited pre-hurricane dataset.

4.3.3.1. Porewater and Extractable Ammonium

Porewater and extractable ammonium, measured at 0-2 and 2-4 cm soil depths, increased throughout the chase period at all sites (Fig. 4.12). However, $\delta^{15}$N enrichment within the sediment ammonium pools was less than 20 $\delta^{15}$N ‰ Air at the Med-OM and Higher-OM sites (Fig. 4.13). Lower-OM ammonium pools enrichment was 52 and 37 $\delta^{15}$N ‰ Air in the 0-2 cm and 2-4 cm sections, respectively. At Lower-OM, $^{15}$N enrichment continued until Hurricane Barry hit and enrichment decreased back to initial levels, with the 2-4 cm section having a slightly higher $\delta^{15}$N signal than the 0-2 cm section. Enrichment in both sections remained above initial $\delta^{15}$N values throughout the entire sampling period. At Med-OM, a very slight $^{15}$N enrichment occurred in 0-2 cm section (15 $\delta^{15}$N ‰ Air), decreasing before Hurricane Barry hit and returning to initial levels. At Higher-OM, slight $\delta^{15}$N enrichment occurred in the 0-2 cm (17 $\delta^{15}$N ‰ Air) and 2-4 cm (18 $\delta^{15}$N ‰ Air) at the 2-week sampling event, returning the initial levels after Hurricane Barry.
Figure 4.12: Total ammonium concentrations within porewater and extractable ammonium pools during the chase experiments (May – Oct 2019). Dashed line indicates Hurricane Barry.
Figure 4.13: Extractable and Porewater $^{15}$NH$_4$ measured during the long-term chase experiments at each site. Dashed line indicates Hurricane Barry.
4.3.3.2. Total Soil Nitrogen

There was significant $^{15}$N enrichment of total nitrogen within the top layers of soil before Hurricane Barry hit the area at all three sites (Fig. 4.14). However, there was large variation between the replicates, especially at Lower-OM. The $^{15}$N signal decreased with depth in pre-hurricane cores. However, after Hurricane Barry, $^{15}$N enrichment was more uniform within the top 5 cm. At all three sites the $\delta^{15}$N signature decreased after the hurricane, but not back to pre-pulse levels. Average 0-5 cm $\delta^{15}$N enrichment peaked at the 2-week, 3-month, and 2-week sampling events for the Lower-OM, Med-OM, and Higher-OM sites, respectively (712, 77, and 38 $\delta^{15}$N ‰ Air). However, $\delta^{15}$N enrichment remained above initial levels throughout the entire chase experiment (Fig. 4.15). Since we only sampled the Higher-OM once after the pulse and before the hurricane, it was hard to predict what trend in $\delta^{15}$N enrichment could have been.

![Figure 4.14: Soil total nitrogen profiles of $\delta^{15}$N ‰ Air enrichment during the long-term chase experiments. Red depth profile indicates the first sampling event after Hurricane Barry.](image-url)
Figure 4.15: Soil total nitrogen (0-5 cm) $\delta^{15}$N enrichment measured during the long-term chase experiments at each site. Grey dashed line indicates Hurricane Barry.

4.3.3.3. Aboveground and Belowground Organic Nitrogen

Vegetation also showed enrichment during the long-term chase experiments, especially the stems (Fig. 4.16). Enrichment of stems at the Lower-OM increased after the pulse, but began to decrease even before Hurricane Barry hits. Enrichment of belowground roots and rhizomes
increased after the pulse and remained relatively constant until Hurricane Barry hit. After the hurricane only dead rhizomes were found and have little to no $\delta^{15}$N enrichment. At Med-OM, *C. esculenta* enrichment decreased before the hurricane hit, while *A. philoxeroides* and root enrichment did not decrease until the hurricane hit. At Higher-OM, the $\delta^{15}$N signal was smaller than the other sites, with similar levels of enrichment in the above and belowground live portions. Both aboveground and belowground live portions maintained enrichment even after the hurricane, although the stems were brown and dying at the time of sampling. Dead stems continued to be present in some cores through the August sampling, with high $^{15}$N enrichment in one mesocosm and no enrichment in the other. Error bars highlight the high variability between each mesocosm, indicating heterogeneous uptake of the $^{15}$N tracer.
Figure 4.16: Vegetation $\delta^{15}$N enrichment measured during the long-term chase experiments at each site. Dashed line indicates Hurricane Barry.

4.3.4. $^{15}$N Mass Balance

Based on the nitrate flux rates of the short-term experiments, approximately 106, 128, and 34 hours of incubation passed before all $^{15}$NO$_3$ tracer was depleted from the water column at Lower-OM, Med-OM, and Higher-OM sites, respectively. Initial sediment uptake rates were calculated as the accumulation of $^{15}$N (by mass) in sediment, stems/leaves, and roots/rhizomes
during the short-term experiments over the estimated time for nitrate depletion in the surface water (Fig. 4.17). Uptake rates were largest on the surface of Lower-OM (649 μg $^{15}$N m$^{-2}$ h$^{-1}$) and relatively small at Med-OM and Higher-OM (95 and 92 μg $^{15}$N m$^{-2}$ h$^{-1}$). Microorganisms within the sediment surface likely contribute to this rate, although these rates are determined indirectly as the total nitrogen minus ammonium and nitrogen within degrading organic material. There is significant variation among mesocosm replicates, especially at Lower-OM.

Initial contribution by aboveground vegetation was smaller at Lower-OM (9.6 μg $^{15}$N m$^{-2}$ h$^{-1}$) and Higher-OM (19.5 μg $^{15}$N m$^{-2}$ h$^{-1}$) than at Med-OM (60.6 μg $^{15}$N m$^{-2}$ h$^{-1}$). There was more aboveground vegetation biomass during the pulse experiments at Med-OM and Higher-OM, which would influence the role of vegetation at these sites compared to Lower-OM.

Figure 4.17. Initial uptake rates within the marsh surface determined from mass accumulation of $^{15}$N over the 2-3-week period of short-term experiments. (a) Vegetation and total sediment uptake rates. (b) Ammonium, degrading organic material, and other nitrogen uptake rates.
Converting nitrate flux, denitrification, and DNRA rates to \( \mu g \, ^{15}N \, m^{-2} \, hr^{-1} \) using initial water depths, we estimated the relative contribution to nitrate removal for each process (Fig. 4.18). Sediment uptake rates were those discussed above, averaged as the accumulation of \(^{15}N\) over the short-term experiments. At Lower-OM, the sum of individual process rates was greater than the total nitrate flux measured in the surface water. On the other hand, at the Med-OM and Higher-OM, 67% and 39% of the added \(^{15}NO_3\) was unaccounted for, indicating either loss through leakage or nitrogen assimilation within the water column. While we were unable to measure \(^{15}N\) in the total organic nitrogen pool, this is a potential sink for the missing tracer. The role of denitrification is also a significant pathway, accounting for 71, 28, and 60% of nitrate fluxes. Extremely low dissolved oxygen concentrations at Higher-OM provide ideal conditions for anaerobic pathways. Sediment loading, nitrate available to benthic communities and vegetation, accounts for 26%, 2%, and <1% of total nitrate uptake at Lower-OM, Med-OM, and Higher-OM, respectively. DNRA accounted for 2%, 2%, and 4% of nitrate uptake, including both porewater and surface rates.
Long term storage of $^{15}$N within the vegetation storage compartments was calculated throughout the chase, including the three months following Hurricane Barry, as a sum of the entire mesocosm footprint (Fig. 4.19). Within the vegetation compartments, the degrading belowground organic material stored a significant portion of the $^{15}$N mass, especially within Lower-OM. Before Hurricane Barry, the aboveground components (stems and leaves) also contributed to $^{15}$N storage. At Higher-OM, we sampled seven days after the hurricane and the $N. lutea$ stems were still present and rigid, but were brown. However, the dead stems still held 7.6
mg $^{15}$N and the live roots and rhizomes stored 10.7 mg $^{15}$N. At the Lower-OM and Med-OM site, we did not sample after Hurricane Barry until August and most of the dead vegetation was gone. Some dead material dropped to the sediment surface and was eventually incorporated into the belowground degrading organic material, but some was assumed to wash away with the tides, exported offshore. At Med-OM, some new vegetation established after the initial destruction from Hurricane Barry and by October 2019, 2.3 mg $^{15}$N was stored within the aboveground compartment.

Figure 4.19. $^{15}$N mass balance within the vegetation components. Red dashed line indicates the occurrence of Hurricane Barry on July 12, 2019.

Enrichment of sediment ammonium pools after the initial pulse indicated diffusion of $^{15}$NH$_4$ after the sediment-water interface (Fig. 4.20). The largest $^{15}$NH$_4$ pools were found in Lower-OM, with a similar distribution between 0-2 and 2-4 cm. There was a slight increase in total (0-4 cm) $^{15}$NH$_4$ between the 2-week (3.7 mg $^{15}$NH$_4$) and 2-month (3.4 mg $^{15}$NH$_4$) sampling.
events. There were much small $^{15}\text{NH}_4$ pools found in the Med-OM (<1 $^{15}\text{NH}_4$) and Higher-OM (<2 mg $^{15}\text{NH}_4$) sites. $^{15}\text{NH}_4$ pools decreased after Hurricane Barry in Lower-OM, but levels were unchanged at Med-OM. At Higher-OM, there was an initial decrease in $^{15}\text{NH}_4$ storage in the 0-2 cm layer and no change in the 2-4 cm layer. Approximately 2 months after the hurricane, $^{15}\text{NH}_4$ pools increased again, possibly indicating decomposition of organic material following plant die-off from Hurricane Barry.

![Graph showing sediment ammonium pools throughout the chase experiments within 0-2 cm and 2-4 cm soil layers. Red dashed line indicates the occurrence of Hurricane Barry on July 12, 2019.](image)

The total soil $^{15}\text{N}$ was also highest at Lower-OM, decreasing significantly after Hurricane Barry, but not back to initial levels (Fig. 4.21). Peak $^{15}\text{N}$ storage occurred at the 2-week sampling event, with a total of 83 mg $^{15}\text{N}$, the greatest amount occurring the in 4-5 cm section, decreasing upward. The Med-OM had a relatively constant soil $^{15}\text{N}$ storage even after Hurricane Barry, peaking at 95 mg $^{15}\text{N}$ at the 6-month sampling event. Higher-OM had the smallest total $^{15}\text{N}$ soil
storage, peaking at the 2-week sampling event with 55 mg $^{15}$N. The total soil nitrogen included the ammonium pools as well as the belowground degrading organic material. A full mass balance of sediment vs. vegetation and alive vs. dead material is shown in Figure 4.22.

Figure 4.21. $^{15}$N mass within soil total nitrogen (top 5 cm) across sites during long-term chase experiments. Grey dashed line indicates the occurrence of Hurricane Barry on July 12, 2019.
Figure 4.22. Full mass balance of $^{15}$N including soil and vegetation compartments. Grey dashed lined indicates the occurrence of Hurricane Barry on July 12, 2019.

### 4.3.5. Disturbance Experiments

Hurricane Barry had a visible effect on the vegetation of WLD, especially in the interior, low elevation areas, where our Lower-OM and Higher-OM sites were located. Most of the *N. lutea* was dead within one week of the hurricane. Some new growth occurred after the hurricane, but not near our experiments. Average 0-5 cm and 5-10 cm bulk densities did not vary significantly over the course of the experiment. We expected some increase in bulk density after Hurricane Barry, based on observations in the field. Within the top 2 cm, there was a slight increase in bulk density at Lower-OM and Higher-OM before and after the hurricane (Fig. 4.23a). These sites were in more exposed areas of the island and could have received sediment during the storm. At Higher-OM, we sampled just after the hurricane and there is an increase in bulk density from 0.23 to 0.38 g cm$^{-3}$. For the other three sites, we sampled just before the hurricane and then again in August. Bulk densities within the top 2 cm at Lower-OM increased...
from 0.86 to 1.06 g cm$^{-3}$. However, the organic matter content of Lower-OM did not decrease after the hurricane, as would be expected with sediment deposition (Fig. 4.23b). This discrepancy could indicate compaction of the soil, creating increased bulk density, but no change in organic matter content. There was no change in bulk density at Med-OM, but there was an increase in organic matter content, from 9.3% to 12.1%, which may be a response to belowground root growth.

Figure 4.23. Changes in (a) soil bulk density and (b) organic matter content after Hurricane Barry.
After the damage caused by Hurricane Barry and the decay of vegetation, we expected to see the decomposition of belowground plant material and the production of $^{15}$NH$_4$ as a result. However, while we did notice that the belowground biomass at our sites were decomposing, we did not detect any $^{15}$NH$_4$ in our core incubation. In all experiments, there was a small, but positive flux of NO$_2$ out of the sediment and a negative flux of NO$_3$ into the sediment (Fig. 4.24a). The smallest NO$_2$ fluxes occurred in the Lower-OM cores, with similar values during the dark and light experiments (3.9 and 0.7 μmoles m$^{-2}$ hr$^{-1}$, respectively). The largest NO$_2$ fluxes occurred in Med-OM during the dark experiment (39.6 μmoles m$^{-2}$ hr$^{-1}$). Med-OM light experiment and Higher-OM dark and light experiment were similar, 17.8, 13.3, and 12.0 μmoles m$^{-2}$ hr$^{-1}$, respectively. Nitrate fluxes were highly variable, but are consistently negative, indicating nitrate influx (Fig. 4.24b). Dark experiments in the Lower-OM cores produced the smallest fluxes (-92 μmoles m$^{-2}$ hr$^{-1}$). Med-OM and Higher-OM dark experiments produced comparable NO$_3$ fluxes (-223 and -210 μmoles m$^{-2}$ hr$^{-1}$, respectively). Lower-OM and Higher-OM light experiments had comparable NO$_3$ fluxes (149 and 145 μmoles m$^{-2}$ hr$^{-1}$), while Med-OM light experiments had the largest NO$_3$ fluxes (308 μmoles m$^{-2}$ hr$^{-1}$).
Figure 4.24. (a) nitrite and (b) nitrate fluxes from the core incubation experiments collected in August 2019. Error bars show standard error for triplicate cores, except at the Med-OM experiments where one core was excluded due to possible contamination.

The effect of light on ammonium fluxes, which switch from negative to positive in Med-OM and Higher-OM cores, suggests light stimulates NH$_4$ uptake (Fig. 4.25a). The Lower-OM cores had negligible NH$_4$ fluxes under dark conditions, but the greatest NH$_4$ uptake rates under light conditions. Lower-OM, Med-OM, and Higher-OM NH$_4$ fluxes during light experiments were -89, -41, and -24 μmoles m$^{-2}$ hr$^{-1}$ and during dark experiments are -15, 173, and 83 μmoles m$^{-2}$ hr$^{-1}$, respectively. Large standard error in the Med-OM cores highlight strong variability in soil characteristics of these sediment cores.

Phosphate fluxes responded to light conditions with an increasing flux of PO$_4$ into the soil under light conditions at all sites (Fig. 4.25b). Fluxes were similar at all three sites for each light condition, increasing on average by 15-20 μmoles m$^{-2}$ hr$^{-1}$ when the lights are turned on.
N\textsubscript{2} production, or denitrification, increased at all sites under light conditions (Fig. 4.26). Under dark conditions, all three sites had similar denitrification rates of 232, 278 and 286 μmoles N m\textsuperscript{-2} hr\textsuperscript{-1} at Lower-OM, Med-OM, and Higher-OM, respectively. Under light conditions, Lower-OM denitrification increased to 264 μmoles m\textsuperscript{-2} hr\textsuperscript{-1}, and Med-OM and Higher-OM increased to 386 and 387 μmoles m\textsuperscript{-2} hr\textsuperscript{-1}, respectively. The similarity of fluxes for the Med-OM and Higher-OM cores suggest that the differences in NO\textsubscript{3} uptake and N\textsubscript{2} production of Med-OM and Higher-OM measured during the field mesocosm experiments are attributable to water column processes, rather than benthic fluxes.
4.4. Discussion

4.4.1. Water Column Processes

In general, we found total nitrate fluxes increase with sediment organic matter content, but the large contribution of water column processes to nitrate uptake suggest that organic matter content is not the driving factor determining these rates. Increases in NO$_x$ fluxes during 2$^{nd}$ pulse, as compared to the 1$^{st}$ pulse, at Lower-OM and Higher-OM highlight the importance of water column conditions to nitrate uptake. At Higher-OM, the increase in nitrate uptake may be due to low nitrate concentrations (< 0.4 μM) and anaerobic conditions (< 0.3 mg L$^{-1}$) that occur after the 1$^{st}$ pulse. Corresponding increases in denitrification during the 2$^{nd}$ pulse suggest that anaerobic respiration is responsible for the increased flux of nitrate. At Lower-OM, nitrate concentrations before the 2$^{nd}$ pulse are still relatively high (40-45 μM) and the water column remains aerobic (6.0-6.8 mg DO L$^{-1}$). Denitrification rates during the 2$^{nd}$ pulse do not increase, indicating increased nitrate removal rates are not attributable to anaerobic respiration. However, there could
still be increased microbial activity caused by the enclosure and the previous pulse, resulting in much larger NO$_x$ uptake rates after the 2$^{nd}$ pulse. Chlorophyll a concentrations increase throughout both pulses at Lower-OM (Appendix C Fig C.3), up to 26 mg L$^{-1}$ during the second pulse. Therefore, water column microbial communities may be stimulated by nitrate addition, generating increased nitrate uptake rates during the 2$^{nd}$ pulse.

Due to the significant contribution of water column processes to nitrate uptake, surface water conditions had a strong impact on the fate of nitrate in these systems. However, water column processes were only temporary nitrate removal processes, especially through assimilation into suspended organic nitrogen. Unlike our mesocosms, residence time of water within coastal deltaic floodplains can be quite short (<1 day) and both unused nitrate and suspended organic nitrogen will move through these systems quickly. Only denitrification or nitrogen burial in the sediment are long-term nitrate removal pathways.

While our field mesocosms incorporate water column, benthic, and vegetation processes, it was difficult to isolate these processes with our dataset. Despite the disturbance caused by Hurricane Barry, the core incubation experiment provided reference benthic fluxes for each site. We used the benthic flux rates from our core incubation experiment to estimate water column contributions to nitrate removal. To normalize our nitrate uptake rates to marsh surface area, we multiplied volumetric fluxes from each experiment by initial water depths within enclosures and average night and day rates when available. Mesocosm nitrate fluxes at Lower-OM, Med-OM, and Higher-OM were 115, 527 and 1557 μmoles m$^{-2}$ hr$^{-1}$ greater than benthic fluxes. These differences were even greater if we use fluxes from the 2$^{nd}$ pulse (1373 and 5124 μmoles m$^{-2}$ hr$^{-1}$ at Lower-OM and Higher-OM, respectively). Therefore, a significant portion of nitrate removal was attributable to water column processes. As mentioned above, increases in chlorophyll a
concentration (Appendix C. Fig. C.3) suggest that phytoplankton, or other microbial communities, may be responsible for a portion of the nitrate uptake at Lower-OM. However, at Higher-OM, chlorophyll a concentrations did not increase throughout the experiment, indicating that increased nitrate fluxes in the mesocosms may be attributable to dissimilatory processes.

While we attempted to separate water column and sediment processes, diffusion across the sediment-water interface was simplified. Denitrification and DNRA rates were calculated using surface water concentrations, but these processes occur within the benthic layers, with the products (N\(_2\) and NH\(_4\)) diffusing back across the sediment water interface. However, it was the prolonged retention of both NO\(_3\) and NH\(_4\) below the sediment surface that we are targeting in this analysis. Those components that did not diffuse back to the water column were what we eventually measured in the plants. Similarly, microbial communities on the sediment surface were important drivers of nitrate uptake and while they contributed to the decrease in surface water nitrate concentrations, their \(^{15}\)N signal was measured in the long-term chase experiments. Water column and sediment surface bacterial and algal biomass can contribute up to 50-100% of surface water nitrate (Hou et al., 2012; Veuger et al., 2007). Changing the amount of time to leave the enclosures in place could impact the outcome of this experiment and change the calculated sediment loading rate. While loading of the tracer into the marsh surface via water column enrichment allowed us to compare water column and sediment processes, direct loading to the sediment is commonly used to increase isotope signal in sediment pools and parametrize long-term budget equations (White and Howes, 1994).

4.4.2. Dissimilatory Processes

Denitrification rates were less dependent on water column processes and increased to some degree with sediment organic matter content in both the mesocosm and benthic flux
experiments. If we normalized the mesocosm denitrification rates by depth, Higher-OM
denitrification rates are 1256 \text{ µmoles m}^{-2} \text{ hr}^{-1} larger in the mesocosm experiments than the core
incubation experiments during the 1\textsuperscript{st} pulse (4919 \text{ µmoles m}^{-2} \text{ hr}^{-1} larger in the 2\textsuperscript{nd} pulse).
Anaerobic conditions likely allow for denitrification to occur within the water column during
both pulses. At both Lower-OM and Med-OM, denitrification rates are 499 and 286 \text{ µmoles m}^{-2}
hr\textsuperscript{-1} higher than benthic fluxes, respectively). Unlike Higher-OM, Lower-OM did not see any
change in denitrification during the 2\textsuperscript{nd} pulse experiment, supporting the idea that denitrification
is limited to the sediment at Lower-OM due to aerobic water column conditions.

Organic carbon concentration, and quality, have a positive correlation with denitrification
rates (Dodla et al., 2008). Therefore, we expected to see an increase in denitrification at our Med
and Higher-OM sites. The effect of organic matter content is consistently found to increase
denitrification rates within sediments, especially freshwater wetlands (Henry and Twilley, 2014).
However, denitrification rates at our sites are high compared to those presented by Henry and
Twilley (2014). Based on high nitrate availability, we expect direct denitrification to be more
significant in this system (Koop-Jakobsen and Giblin, 2010; Seitzinger et al., 2007)
Salt marshes under high “fertilized” nitrate conditions exhibit higher rates of
denitrification (294 \text{ µmoles m}^{-2} \text{ hr}^{-1}) compared marshes under low nitrate (< 7 \text{ µM} 9.6 \text{ µmoles m}^{-2}
\text{ hr}^{-1}) conditions (Koop-Jakobsen and Giblin, 2010). As a distributary of the Mississippi River,
Wax Lake Outlet delivers large nitrate loads to the WLD, regularly above 40 \text{ µM}. Therefore, we
expect higher rates of denitrification at WLD, as compared to non-fertilized marshes, and we
don’t expect the nitrate additions to have a large impact on denitrification rates. However, due to
near-depletion nitrate concentrations after 48 hours of deployment, Higher-OM denitrification
rates increase significantly during the 2\textsuperscript{nd} pulse.
DNRA has been shown to increase with salinity (An and Gardner, 2002; Gardner et al., 2006), especially relative to denitrification, due to the sulfate inhibition of denitrification and competition for organic electron donors with sulfate-reducing bacteria which have a capacity to perform DNRA (Brunet and Garcia-Gil, 1996; Rysgaard et al., 1996). Our results support this trend with very low DNRA rates in our sites. In addition to denitrification and DNRA, another potential dissimilatory process is anaerobic ammonium oxidation (anammox), which has a significant contribution (up to 50%) to N\textsubscript{2} production in marine sediments (Dalsgaard et al., 2005). However, less is known about the contribution of anammox in shallow, freshwater marshes (Zhu et al., 2010).

4.4.3. Sediment Assimilation

We found high \textsuperscript{15}N uptake rates to the marsh surface sediment at the Lower-OM, but lower rates at the Med-OM and Higher-OM, which indicates 1) high assimilation rates in the water column of those sites and limited diffusion through the sediment-water interface and/or 2) competition by denitrifying bacteria in the surface sediments. However, by leaving the mesocosms in place for 2-3 weeks, we may underestimate initial loading to the surface sediments. Remineralization of organic nitrogen on the surface may occur within our inoculation period, a process our experimental design cannot quantify (Erler et al., 2010). Eyre et al. (2016) found that the 50% percent of \textsuperscript{15}N incorporated in the surface sediments of a subtidal estuary was lost over 6 days. Throughout the 33-day experiment, 20.8\%, 16.5\%, and 20.7\% of the sediment organic nitrogen pool was lost through remineralization, nitrification, and denitrification. At Lower-OM, there is a decrease in total \textsuperscript{15}N pools, even before Hurricane Barry, which may indicate remineralization of surface organic nitrogen, which diffuses to the surface water as \textsuperscript{15}NH\textsubscript{4}, or undergoes coupled nitrification-denitrification. We did not measure surface water
components during the long-term experiments. Our disturbance core incubations indicate potential mineralization of sediment organic nitrogen with ammonium efflux under dark conditions at the Med-OM and Higher-OM sites. However, no ammonium release was measured at Lower-OM or the under light conditions at any site.

In cobble and sand streams, assimilation by algal and microbial biomass accounted for 88% of nitrate uptake, of which 0.3%, 5.4%, and 0.1% was remineralized to ammonium, nitrified to nitrate, and denitrification to dinitrogen gas (O’Brien et al., 2012). In shallow, nitrogen-limited, saline bay, assimilation accounted for 50-70% (Hou et al. 2012) and 50-100% (Veuger et al., 2007) of nitrate. In a tidal, freshwater marsh, assimilation by microbes on litter and surface sediments was found to be more important than plant uptake (41-83%; Gribsholt et al., 2007). Similarly, in a freshwater treatment wetland, 9-40% of added ammonium was assimilated into plant and microbial biomass after 1-2 days, but a significant portion was remineralized with 31% and 7.4% remaining in the sediment and plants over 157 days (Erler et al., 2010).

While microbial nitrogen is recycled quickly, nitrogen within macrophytes has a longer retention time, generally increasing throughout the growing season. Freshwater wetland vegetation, especially submerged aquatic vegetation, is a significant sink for surface water nitrogen (Tan et al., 2013). While Hurricane Barry shortened the growing season at WLD, we did see a transition of $^{15}$N stored in the soil to vegetation at the Med-OM and Higher-OM sites. Unfortunately, we only measured $^{15}$NH$_4$ pools within the top 4 cm of sediment, despite rooting depths down to 30-40 cm and hypothesize that we are missing the $^{15}$NH$_4$ at deeper depths. This may explain why total $^{15}$N pools increase from the 2$^{nd}$ to 3$^{rd}$ sampling events at Med-OM and Higher-OM. With the aboveground vegetation die-off and slow decomposition of belowground roots and rhizomes, we expect that a significant portion of the $^{15}$N stored in plants is exported.
offshore. However, a certain percent is likely incorporated into the dead organic material within
the sediment profile, although we did not detect that in our experiments.

Over a full growing season, loss of ammonium (injected directly into the soil) is greater
in freshwater (20%; 18% is maintained in aboveground vegetation and 62% in roots and soil;
Delaune et al., 1986) than in salt marshes (7%; 28% is maintained in aboveground vegetation
and 65% in roots and soil; Buresh et al., 1981). Most of the loss is attributed to export or
remineralization of aboveground plant material. However, salt marshes are dominated by
perennial graminoids with extensive rhizome complexes, and may have lower export of
aboveground plant material depending on marsh hydrology. Extending experiments past one
growing season (DeLaune et al., 1983) would improve our budgets of nitrogen loss through plant
export, which is currently understudied in coastal deltaic floodplains

The aim of this project was to quantify the fate of nitrate in coastal deltaic floodplains.
Based on the results, we estimate that more than 37-66% of surface water nitrate remains in the
water column as nitrate or suspended organic nitrogen, while 28-71% is lost to the atmosphere as
N₂, and 1-26% is stored within the sediment, benthic microbes, and macrophytes pools at least
through the growing season. Disturbances, such as hurricanes, can increase export of organic
nitrogen, but may also increase nitrogen storage within sediments through the deposition of
sediment on the marsh surface. More data on sediment transport is needed to understand the role
of hurricanes in sediment nitrogen cycling.

4.2.4. Fate of Nitrate at Wax Lake Delta

The fate of nitrate at WLD depends on several factors, some of which our experimental
design did not incorporate, including residence time and temperature (Dettmann, 2001). Most
importantly, these factors control the amount of time microbially-mediated processes interact
with and process nitrate in the surface water. Our experimental design greatly increased the residence time of water within these wetland systems. We left enclosures in place for 2-3 weeks to maximize transfer of the tracer to the marsh surface. Tidal fluctuations created artificially high exchange, pumping water across the sediment-water interface at greater rates than a coastal system typically undergoes except under high wind and wave conditions (Risgaard-Petersen et al., 1999). Despite this, there was limited nitrate loading to the marsh surface, indicating the importance of water column processes.

Breton Sound estuary, which receives Mississippi River water through the Caernarvon freshwater diversion, removes 88-95\% of surface water nitrate through assimilation and denitrification (Lane et al., 1999). These results indicate that Breton Sound wetlands can tolerate large nitrate loads without detrimental impacts. Unlike Breton Sound, WLD’s low connectivity between channels and floodplains reduces the impact of wetland soils on high riverine nitrate concentrations. Previous estimates suggest a nitrate removal efficiency of 41-47\% in the Atchafalaya Coastal Basin (66 to 172 g N m$^{-2}$ yr$^{-1}$), which includes the forested wetlands, deltaic wetlands, and coastal estuary (Lane et al., 2002). However, the role of WLD is quite small, but should increase as the total delta area increases.

4.5. Conclusions

Overall, the fate of nitrate depends on the biogeochemical process measured in this study as well environmental and physical factors related to the hydrology of WLD. First, increased residence time and hydrological regime can lead to greater nitrate removal through surface water processes and diffusion into the porewater (Martin and Reddy, 1997; Tanner and Kadlec, 2013; White et al., 2019). Second, water temperature plays a role in most microbially-mediated processes. Water temperature is a function of meteorological conditions and residence time.
Third, while we found linear nitrate uptake rates, many nitrate processes are concentration dependent and nitrate loading to the ecosystem plays a significant role in nitrate removal. In coastal deltaic floodplains, the river is the source of nitrate and connectivity between channels and floodplains determines the delivery of nitrate. Connectivity varies temporally with changes in river stage, and spatially, based on floodplain geomorphology and vegetation structure. At WLD, as little as 14% of channel water enters the floodplains. During high river discharge, while larger volumes enter the floodplains, the amount of time spend within the floodplains is greatly reduced.

While this study has increased our understanding of nitrate removal pathways at WLD, incorporating these rates into hydrodynamic models will allow us to simulate nitrate uptake and denitrification across the entire delta. WLD can be used as an analog for proposed river sediment diversions from the Mississippi River as part of the Louisiana Coastal Master Plan (CPRA, 2017). The development of a full nitrogen uptake model, coupled with hydrodynamics, would be a powerful tool for studying the role of coastal deltaic floodplains in nitrate removal and optimizing restoration efforts along the coast of Louisiana.
Chapter 5. Modeling the Fate of Nitrate in a Coastal Deltaic Floodplain

5.1. Introduction

Large nitrogen loads in the Mississippi River are a result of agricultural and urban runoff and contribute to eutrophication and hypoxia in the Gulf of Mexico (Burkart and James, 1999; Goolsby and Battaglin, 2001; Rabalais et al., 2002; Turner and Rabalais, 1991). While demands on food production are increasing as human populations grow, solutions for nitrogen loading to the Gulf of Mexico include improved farm practices as well as use of wetlands for nitrogen removal (Donner and Kucharik, 2008; Galloway et al., 2008; Mitsch et al., 2005a, 2001). Expansion of riparian wetlands along the tributaries and distributaries of the Mississippi River as well as the use of river diversions for removal of nitrogen are examples of previously proposed strategies (Mitsch et al., 2001). Wetlands within the Mississippi River Basin retain 39-46 gN m$^{-2}$ yr$^{-1}$ and have the capacity to remove 40% of the nitrogen discharging to the Gulf of Mexico (Mitsch et al., 2005).

Coastal deltaic floodplains, at the boundary between terrestrial and aquatic and river and tide-dominated ecosystems, have been shown to exhibit complex hydrology and biogeochemistry, with hourly, seasonal, and annual variability. These wetland ecosystems are recognized for providing many ecosystem services, including improving water quality by retaining nitrogen produced from urban and agricultural runoff. Through exchange between channels and floodplains, large nitrogen loads are delivered to wetlands, which can be a sink for nitrate and other nitrogen compounds.

While water quality models are an important tool for assessing nitrate loading conditions, the calibration and validation required often limit the ability of models. The difficulty of incorporating reaction rates across large temporal and spatial scales can lead to significant errors.
because these are site and time specific. Often small areas are responsible for a significant amount of total nitrate removal, but applying these rates to larger regions may overestimate nitrate removal (Groffman et al., 2009). While temporal data are becoming more accessible with the installation of more continuous monitoring stations, high-resolution spatial data are still difficult to find. For example, hydrologic and biogeochemical factors can contribute equally to nitrogen loss through denitrification (Ensign et al., 2013). Therefore, environmental conditions such as water temperature, inundation duration and frequency, water residence time, surface topography, and soil organic carbon content are important variables to incorporate into a nitrogen model (Appelboom et al., 2010; Bachand and Horne, 1999; Dettmann, 2001; Dodla et al., 2008; Frei et al., 2012; Tanner and Kadlec, 2013). While the first four factors can be predicted through modeling and remote sensing, spatially explicit patterns of carbon substrate are less easily obtained.

There are currently three common approaches for modeling nitrate removal: 1) coupling complex 2D hydrodynamic models with simple nitrate uptake models (Hiatt et al., 2018; Messer et al., 2017), 2) simple 1D hydrology models paired with multiple compartment, process-specific nitrogen models (Kadlec and Hammer, 1988), and 3) a full 3D incorporation of hydrodynamic and biogeochemical processes (Grégoire and Beckers, 2004; Hantush et al., 2013; Smits and van Beek, 2013). While a full nitrate model should include water column and surface process, the logistics of such a complex model, including the error associated with each process, led us to a simpler nitrate uptake model, incorporating net nitrate fluxes and denitrification only. We focus on the first approach, using a 2D hydrodynamic model to explore methods for applying general nitrate uptake rates across an entire system. While capturing spatial and temporal
biogeochemical hotspots is quite difficult (Groffman et al., 2009), we aim to capture general
trends and overall nitrate removal capacity of a coastal deltaic floodplain.

5.2. Methods

5.2.1. Study Site

Our study site, Wax Lake Delta (WLD), is a newly emergent, actively prograding delta in
southeastern Louisiana (Fig. 5.1a). WLD formed after the dredging of the Wax Lake Outlet by
the US Army Corps of Engineers to reduce flooding in Morgan City, Louisiana in 1944. After a
large flood in 1973, WLD became subaerial and is estimated to be growing at 1-2 km² hr⁻¹ (Kim
et al., 2009) WLD receives approximately 11% of the Mississippi River discharge and nitrate
loading to WLD, which varies with season and discharge, as high as 100μM. A significant
portion of the flow through the primary channels of WLD interacts with the floodplains (up to
40%, Chapter 2). However, this interaction is concentrated at the delta front, where residence
time is shorter, sediment organic matter content is lower, and very little vegetation establishes.
Figure 5.1. Study site, Wax Lake Delta, located in southeastern Louisiana (A) and the model domain, classified into hydrogeomorphic zones by sediment bed elevation (B-C).

5.2.2. Delft3D Model Description

Delft3D modeling suite is chosen for this project because of its applicability to shallow water systems and for capacity for coupling water quality and hydrodynamic simulations. Delft3D-Water Quality (Delft3D-WAQ) simulates mass transfer of a variety of substances. Delft3D-WAQ can be coupled with the hydrodynamic output from Delft3D-FLOW (Deltares, 2016), with options for both spatial aggregation and longer time steps. Delft3D-WAQ solves the advection-diffusion equations, along with specific reaction processes defined by partial differential equations. Unlike Delft3D-FLOW, Delft3D-WAQ uses a finite volume method,
which evaluates mass exchange across the surface of a volume. Delft3D-WAQ solves the advection diffusion equations:

\[
M_i^{t+\Delta} = M_i^t + \Delta t * \left( \frac{\Delta M}{\Delta t} \right)_{tr} + \Delta t * \left( \frac{\Delta M}{\Delta t} \right)_{p} + \Delta t * \left( \frac{\Delta M}{\Delta t} \right)_{s}
\]

Eq 5.1

Model domain includes WLD, with open boundaries in the Wax Lake Outlet and offshore (Fig. 5.1b). Details of the hydrodynamic model and calibration are detailed in Chapter 2. For each model, rates of nitrate removal are applied only across the major deltaic islands of WLD. Each model incorporates nitrate flux rates from different sources, under 1st order or 0th order reaction rates. These fluxes incorporate either total nitrate flux or only denitrification. Denitrification is the major process in estuaries, accounting for 69-75% of net nitrate removal (Dettman, 2001). While assimilation, dissimilatory nitrate reduction to ammonium, diffusion and loading to sediment and vegetation, and any other process controlling the uptake of surface water nitrate, contribute to total nitrate, they are not modeled individually.

To apply field rates, we classified WLD into hydrogeomorphic (HGM) and chronosequence zones. HGM were identified across WLD based on elevation from USGS National Elevation Dataset (Lidar derived) Digital Elevation Model from 2012, where zones are distinguished using the local tidal datum (Fig. 5.1.b-c). Classification of chronosequence zones is modified from Bevington and Twilley (2018) to produce a quantitative approach. Elevations along cross-sectional transects (every 10m) are averaged and chronosequence zone classifications are applied based on the average elevation and corresponding HGM zone (Fig. 5.2). In general, older chronosequences (I-II) are dominated by supratidal and high intertidal zones, while younger chronosequences (III-IV) are dominated by low intertidal and subtidal zones. Using this approach, we can classify each island objectively.
A constant upstream nitrate concentration of 100 μM in the Wax Lake Outlet was consistent with high nitrate loading measured during summer months. Through the 2-month simulation period, a total 29,000 Mt N was loaded to the system. Nitrate removal efficiency was calculated as amount (by mass) of nitrate-nitrogen retained, calculated as

$$\text{N Retention} = 100 \times \frac{N_{\text{loaded}} - N_{\text{exported}}}{N_{\text{loaded}}}$$

Eq. 5.2

5.2.2.1. Models – Field Mesocosm

The first three models incorporated field measured rates for denitrification, water column assimilation, and sediment assimilation based on experiments done at WLD in spring/summer 2019 (Chapter 4). Experiments were run in field mesocosms using a $^{15}$NO$_3$ surface water tracer
to monitor the loss and transformation of nitrate through denitrification, dissimilatory nitrate reduction to ammonium, and water column assimilation. Sediment assimilation encompasses plant and microbial uptake within the surface sediments. The remaining nitrate loss is attributed to water column assimilation, which accounts for 30-80% of nitrate fluxes. Only three sites were included in these experiments, representing low, medium, and higher sediment organic matter content as well as subtidal and high intertidal HGM zones. Rates from Med-OM are applied to all supratidal areas of WLD, dominated by *Colocasia esculenta* and *Salix nigra*, with lower inundation frequency and higher soil porosity due to roots and rhizomes. Despite similar elevations, vegetation types, and chronosequence zones, the Lower-OM and Higher-OM sites have distinct soil properties and nitrate uptake rates. Therefore, we applied Lower-OM rates to low intertidal and subtidal zones in chronosequence III and IV, where we expect lower sediment organic matter content. Higher-OM rates are applied to low intertidal and subtidal zones in chronosequence I and II, where we expect higher sediment organic matter content. Rates from Lower-OM and Higher-OM are averaged and applied across all low intertidal and subtidal areas of WLD, dominated by *Nelumbo lutea* and submerged aquatic vegetation.

Model 1 rates were derived from total nitrate uptake, assuming a first-order function (1 day⁻¹). A temperature coefficient of 1.07 is applied to account for the influence of temperature on these microbially mediated processes. However, we simulated summer months (April - May 2015), so temperatures in the floodplains rarely fell below 18°C. Model 2 rates were derived from zero-order rates from change in concentration of surface water nitrate over several hours (gN m⁻³ d⁻¹). Model 3 rates were derived specifically from denitrification rates calculated on an area basis (gN m⁻² d⁻¹). Final rates for models 1 – 3 are listed in Table 5.1 and spatial classifications are shown in Figure 5.3.
Table 5.1. Reaction rates for models 1-3 based on rates measured in field mesocosm experiments (Chapter 4).

<table>
<thead>
<tr>
<th>Hydrogeomorphic and Chronosequence Zone</th>
<th>Model 1 Nitrate uptake (d(^{-1}))</th>
<th>Model 2 Nitrate Uptake (gN m(^{-3}) d(^{-1}))</th>
<th>Model 3 Denitrification (gN m(^{-2}) d(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Supratidal and High Intertidal I - IV</td>
<td>0.15</td>
<td>0.14</td>
<td>0.162</td>
</tr>
<tr>
<td>Subtidal and Low Intertidal I - II</td>
<td>0.49</td>
<td>0.85</td>
<td>0.058</td>
</tr>
<tr>
<td>Subtidal and Lower Intertidal III - IV</td>
<td>1.99</td>
<td>0.73</td>
<td>0.496</td>
</tr>
</tbody>
</table>

Figure 5.3. Classifications based on field mesocosm experiments with rates from these experiments applied to Model 1, 2, and 3 accordingly.
5.2.2.2. Literature Denitrification Rates

Model 4 used HGM and chronosequence zones classifications to apply denitrification rates from determined by Li et al. (in press) using sediment core incubation studies and the N₂/Ar analytical methods during the summer 2018. Nine sites represented nine HGM and chronosequence zone combinations (Table 5.2), where high intertidal and low intertidal were combined (Fig. 5.4). Missing combinations are Supra-III, Supra IV, and Sub-I, which represent a small area of the delta. Rates for these areas were combined with Supra-II, Supra-II, and Sub-II, respectively. Denitrification rates (g N m⁻² day⁻¹) represent benthic processes and do not incorporate the role of water column processes.

Table 5.2. N₂ production rates from Li et al. (2020) measured across Mike Island of WLD during the summer (May-August) 2018 used in Model 4

<table>
<thead>
<tr>
<th>HGM - Chronosequence</th>
<th>N₂ Production Rate (g N m⁻² d⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Supratidal - I</td>
<td>0.394</td>
</tr>
<tr>
<td>Supratidal - II/III/IV</td>
<td>0.645</td>
</tr>
<tr>
<td>Intertidal - I</td>
<td>0.363</td>
</tr>
<tr>
<td>Intertidal - II</td>
<td>0.181</td>
</tr>
<tr>
<td>Intertidal - III</td>
<td>0.243</td>
</tr>
<tr>
<td>Intertidal - IV</td>
<td>0.140</td>
</tr>
<tr>
<td>Subtidal - I/II</td>
<td>0.179</td>
</tr>
<tr>
<td>Subtidal - III</td>
<td>0.234</td>
</tr>
<tr>
<td>Subtidal - IV</td>
<td>0.140</td>
</tr>
</tbody>
</table>
Figure 5.4. Classifications based on hydrogeomorphic and chronosequence zones used to applied denitrification rates from Li et al. 2020 for Model 4.

5.2.2.3. Rates Modeled by Organic Matter Content

Models 5a-d scaled denitrification rates to the delta using relationships between sediment organic matter content and N\textsubscript{2} production rates. This relationship is well established, but likely varies by system, especially between marsh type (saline vs. fresh), climate (temperate vs subtropical), and season (winter vs. summer). Therefore, we used the relationship developed by Li et al. 2020 (r\textsuperscript{2} = 0.55, p < 0.001), which was developed using flow-through core incubation experiments across WLD during summer 2017. Four predictive maps of sediment organic matter content at WLD (Chapter 3) were used to develop maps of denitrification (Fig 5.5) according to

$$\text{N}_2\text{-N Flux (\mu moles N m}^{-2}\text{ hr}^{-1}) = 132.0 + 36.3 \times \text{Organic Matter Content (\%) Eq 5.2.}$$
Figure 5.5. Modeled denitrification rates based on organic matter content predictions (Chapter 3) using relationship between N$_2$ production and sediment organic matter from Li et al. 2020. Rates for Model 5a-d are shown in A-D, respectively.

5.3. Results and Discussion

Average nitrate concentrations across the delta floodplains remained above 75 μM for all simulations (Fig. 5.6). Snapshots of each model for June 1, 2020 highlight reduced concentrations and biogeochemical hotspots within floodplains and spatial variation among models (Fig. 5.7). Model 1, first order nitrate uptake, produced the greatest reduction in nitrate concentration within the islands, with concentrations consistently 10 μM less than the other models. In general, the models with field-determined rates were very similar (Models 2, 3, and 4) and models with rates extrapolated from organic matter content (Models 5a-d) were very similar. From the delta front to the delta apex, average channel nitrate concentrations decreased,
indicating removal of nitrate as water flows to the Gulf of Mexico. However, average channel nitrate concentrations remained above 90 μM in all simulations (Fig. 5.8).

Figure 5.6. Average floodplain nitrate concentrations for each model simulation.
Figure 5.7. Snapshots of nitrate concentration from all models on June 1, 2020 at 12:00pm.
Figure 5.8. Average channel nitrate concentration from the delta apex to the delta front in each model simulation.

While model results varied depending on the reaction rates applied, the total contribution of the delta to nitrate retention was similar in all models. Of the 29000 Mt N loaded to the system throughout the simulation, 4.1%, 2.6%, 3.4%, and 2.4 - 2.5% were removed according to Model 1-2, Model 3, Model 4, and Models 5a-d, respectively. Therefore, the nitrate removal potential is very low for the floodplains of WLD during summer months. However, during winter and fall when river discharge is lower and residence time is longer, WLD may have a greater capacity for removing nitrate.

The difference between first order and zero order models highlighted the importance of applying appropriate rates within a water quality model. As the first order model (Model 1) was concentration and temperature dependent, we expect these to be more realistic, especially when the simulation period extends into colder months. However, field observations often do not test
the effect of temperature and nitrate concentration, making first order reaction rates difficult to obtain. Models 5a-d, which were built using the relationship between sediment organic matter and N₂ production, have the lowest estimates for nitrate removal. The sediment organic matter maps were derived using field data within the top 10 cm of sediment. However, much of the denitrification occurs within even shallower sediment depths. Therefore, Models 5a-d may underestimate denitrification rates simply because they are based on underestimates of organic matter content.

With so many factors contributing to the nitrate removal potential of an ecosystem, it’s important to consider the relative importance of hydrological or biogeochemical processes. Significant knowledge can be borrowed from constructed and treatment wetland research and application, especially regarding the role of vegetation (Bachand and Horne, 1999; Kadlec, 2012; Thullen et al., 2005), hydrology (Baker, 1998; Fink and Mitsch, 2007; Kadlec, 2012), and soil chemistry (Kadlec and Reddy, 2001; Lin et al., 2007). Considerations of both physical and biological factors should be made when designing coastal restoration or managing natural systems to optimize ecosystem services.

5.4. Conclusion

This preliminary modelling effort implies a limited potential of WLD to reduce nitrate loading to the Gulf of Mexico. However, WLD is a young delta, growing at 1-2 km² yr⁻¹ (Kim et al., 2009) and its capacity for nitrate removal is expected to increase over time. Within the larger system, WLD represents only a small portion of the Atchafalaya Delta Bay. The role of forested wetlands north of WLD, the Atchafalaya Delta and Fourleague Bay to the east of WLD, and the coastal estuary of the Atchafalaya Bay all contribute to the total nitrate removal capacity of coastal ecosystems. Lane et al. (2009) estimated 41-47% decrease in nitrate concentration from
the Atchafalaya river to the Gulf of Mexico. Similarly, Perez et al. (2010) estimated that 51% of nitrate entering Fourleague Bay, with an average residence time of 7 days, was removed. The overall nitrate removal capacity of coastal ecosystems in Louisiana is significant with physical and biological factors contributing to the spatial and temporal variation in rates of removal and retention efficiency.
Chapter 6. Conclusions

Through the chapters of this dissertation, we have explored physical and biological factors controlling the fate of nitrate in coastal deltaic systems. Chapter 1 identified the hydrological and ecological processes that make coastal deltaic floodplain unique and important systems. Hydrodynamic models developed in Chapter 2 quantified the exchange between channels and floodplains and the impact of river and tide forcings on water residence time. Soil and vegetation surveys in Chapter 3 identified potential biogeochemical hotspots where vegetation resistance and sediment organic matter create optimal conditions for nitrate uptake and denitrification. Field mesocosm experiments in Chapter 4 identified important nitrogen pathways that determine the fate of nitrate at WLD. Finally, Chapter 5 applied different rates of nitrate uptake and denitrification to a simple water quality model of WLD. WLD’s role in nitrate removal is limited, because of limited exchange and a small areal footprint. However, WLD is just one piece the Atchafalaya Coastal Basin and the alluvial floodplains and coastal estuaries produce a significant overall impact on nitrate removal. As WLD progrades and area increases, it’s role in nitrate removal will increase as well. This research emphasizes the strengths and weakness of small and large-scale research and the need for more synthesis of available data in producing landscape-scale estimate of nitrate removal potential. WLD is a natural laboratory, from which coastal managers and state agencies can learn much about ecosystem adaption and restoration design.

Future Considerations

Throughout this project, we’ve learned many lessons in strategizing field, laboratory, and numerical modeling tools for exploring research questions. Appendix C gives a brief description of mesocosm trial and error we experience throughout this research with explanation of
techniques we tried. In addition to those perspectives, there are several considerations of how this research could move forward in the future. While we attempted to develop a nitrogen budget for Wax Lake Delta that incorporated vegetation, sediment, and water column processes, Hurricane Barry limited the extent of this budget. A nitrogen budget that includes seasonal changes included export of organic and inorganic nitrogen could provide environmental managers and state agencies more concert numbers to use in long-term models. In my opinion, the most valuable results for landscape design and management would be short-term nitrate removal capacity (%) and long-term storage and burial (%). The former would include both permanent los through denitrification as well as short-term loss through DNRA and assimilation in the water column and sediment. Long-term storage and burial would incorporate the net assimilation of nitrogen delivered through rivers, accounting for uptake during the growing season and export during winter. While significantly simplified from what we measured in our mesocosm experiments, these values would be enough to estimate the impact of WLD on nitrate retention. However, I believe it’s still important to incorporate spatial variability in these values as they can vary significantly across landscapes, especially across different wetland ecosystems. Our models indicate that general estimates by hydrogeomorphic zone or vegetation type are sufficient for estimating nitrate removal. The role of hydrodynamics may in fact be more important.

One future research goal is to test the impact of delta geomorphology on nitrate removal capacity. Specifically, we are interested in understanding the impact of channel size and number, delta shape and orientation, and relative impact of hydrologic forcings. Ideally, this would involve developing hypothetical deltaic landscapes (such as those in Edmonds and Slingerland, 2010), applying our nitrate uptake rates, either averaged or classified by elevation, and testing
the impact of different deltas on nitrate removal potential. Building deltas would allow us to control relative river, wave, and tide energy to create a spectrum of delta shapes (Fig 6.1; Galloway, 1975). One interesting idea we could explore by building deltas is the development of sediment organic matter through biological feedbacks and the corresponding nitrate uptake rates.

![Figure 6.1. Schematic diagram illustrating the threefold division of deltas into fluvial-dominated, wave-dominated and tide-dominated types. Source: Galloway, 1975](image)

Another future research goal is to put Wax Lake Delta in the context of the Atchafalaya River Basin, which includes alluvial floodplains, the Atchafalaya Delta, Fourleague Bay, and the Atchafalaya Bay estuary. Varying levels of nitrogen removal have been reported in the literature, making it difficult to identify the full capacity of this system (BryantMason et al., 2013; Lane et al., 2002; Perez et al., 2011). Potential denitrification rates in cypress swamps in the Atchafalaya River Basin are 163 and 228 μmoles N m⁻² hr⁻¹ at 22 and 30°C, respectively (Lindau et al.,
2008). However, across habitats in the basin (cypress swamps, bottomland hardwood forests, and lakes), significant spatial variation exists (Scaroni et al., 2010). Fourleague Bay is less well studied, but an estimated 50% of nitrate entering the system is removed via denitrification (Smith et al., 1985). By incorporating literature rates for nitrate uptake in salt marsh, bay sediment, forested floodplains, and freshwater deltas into a larger scale Atchafalaya model, we could predict the overall nitrate removal capacity of the system. We expect that the low removal capacity of WLD predicted by our model (Chapter 5) is related to the size of the delta, which is small compared to the entire Atchafalaya River Basin. Thus, we expect the other ecosystems within the basin to have greater removal capacity, as Lane et al. (2002) have shown that 41-47% of riverine nitrate is removed before reaching the Gulf of Mexico.

A final future research goal is to take advantage of our model to explore the influence of timing on the downstream effect of opening river sediment diversions. First, we would expand our model to all seasons, to incorporate differences in vegetation abundance, water temperature, and river discharge. We expect nitrate removal capacity (%) to increase with lower river discharge (and longer residence times), but to decrease when water temperatures are very low. Second, by adding a downstream coastal phytoplankton community, we could simulate high nitrate export through WLD and compare the phytoplankton response under different timing scenarios. These scenarios could include different seasons as well as varying frequency and duration of floods. There are numerous avenues to explore with this model and with WLD as a natural laboratory for future field and modeling efforts.
Appendix A. Supplementary Material for Chapter 2

Figure A.1. Map of the study area with insert A showing the Louisiana coast and insert B showing satellite image of Wax Lake Delta. Data sources for the Delft3D model are indicated with squares, parentheses identify source.
Table A.1. Summary of data sources for boundary conditions and other model parameters.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Data Source</th>
<th>Station Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>River Discharge</td>
<td>USGS</td>
<td>Wax Lake Outlet at Calumet (29° 41’ 52” N, 91° 22’ 22” W)</td>
</tr>
<tr>
<td>Downstream Water Level</td>
<td>NOAA</td>
<td>Atchafalaya Delta at Amerada Pass (29° 26.9’ N, 91° 20.3’ W)</td>
</tr>
<tr>
<td>River Water Temperature</td>
<td>USGS</td>
<td>Atchafalaya River at Morgan City (29° 41’ 33.4”N, 91° 12’ 42.6” W)</td>
</tr>
<tr>
<td>Downstream Water Temperature</td>
<td>NOAA</td>
<td>Atchafalaya Delta at Amerada Pass (29° 26.9’ N, 91° 20.3’ W)</td>
</tr>
<tr>
<td>Wind Speed and Direction</td>
<td>NOAA</td>
<td>Atchafalaya Delta at Amerada Pass (29° 26.9’ N, 91° 20.3’ W)</td>
</tr>
<tr>
<td>Relative Humidity, Cloud Cover, Air Temperature</td>
<td>IEM</td>
<td>Patterson Airport (29° 42’ 34.38”N, 91° 20’ 20.33”W)</td>
</tr>
</tbody>
</table>
Figure A.2. Boundary conditions and model parameters during the study period of January – June 2015. A) Discharge in the Wax Lake Outlet, B) water Level in the Atchafalaya Delta, C) water temperature in the Atchafalaya River, D) water temperature in the Atchafalaya Delta, E) air temperature, F) relative humidity, and G) cloud cover at Patterson Airport, and H) Wind speed and direction in the Atchafalaya Delta
Figure A.3. A) Location of Delta Dynamics Observatory on Mike Island, B) Instruments on each platform, and C) Correlation plots from calibration of the Delft3D Model
Figure A.4. Channel-floodplain exchange using method 1, separated into flow in (grey) and flow out (black), for each radial section of the delta (approximately every 1000m from the delta apex). Each section is treated as a control volume and values are shown in two-hour increments.
Figure A.5. Six-month average of channel-floodplain exchange (m³/s). Positive (blue) indicates flow into the floodplains and negative (red) indicates flow out of the floodplains. Highlights areas of greatest exchange and increases in exchange beginning in March.
Appendix B. Supplementary Material for Chapter 3

Figure B.1. Trends in stem height for A) *Colocasia esculenta* and B) *Nelumbo lutea* across hydrogeomorphic zone and age chronosequence zone for three sampling months (March, June, and September/October).

Figure B.2. Trends in biomass for A) *Colocasia esculenta* and B) *Nelumbo lutea* across hydrogeomorphic zone and age chronosequence zone for three sampling months (March, June, and September/October).
Figure B.3. Trends in stem density for A) *Colocasia esculenta* and B) *Nelumbo lutea* across hydrogeomorphic zone and age chronosequence zone for three sampling months (March, June, and September/October).

Figure B.4. Trends in plant volume for A) *Colocasia esculenta* and B) *Nelumbo lutea* across hydrogeomorphic zone and age chronosequence zone for three sampling months (March, June, and September/October).
Appendix C. Supplementary Material for Chapter 4

To test the $^{15}$NO$_3$ tracer incubation within field mesocosms, we performed several pilot studies. In 2017, we collaborated with Laurel Larsen from UC Berkeley who install two 10m x 1m field flumes at WLD (Fig. D.1.). These experiments were designed to create 5-10 cm s$^{-1}$ flow through *Nelumbo lutea* and *Colocasia esculenta*, which have distinct stem densities and structures. However, due to the long uptake times for nitrate, we did not expect to detect a change in nitrate concentration from the upstream to the downstream end of the flume. Instead, we closed both ends of the flume and created no-flow chambers. We performed two experiments (April and June) at the *C. esculenta* site, but only one experiment (April) at the *N. lutea* site, due to die off of plants after a storm. From these experiments, we measured total nitrate uptake rates of 3700 and 3900 μmoles m$^{-3}$ hr$^{-1}$ within the *C. esculenta* and *N. lutea* flumes respectively. However, there were often problems with sealing the ends of the chamber, as it was not originally designed or constructed to be sealed.

Figure C.1. Field flumes designed by Laurel Larsen. Photo shows the *Colocasia esculenta* site.
In 2018, we designed and constructed a smaller, portable field flume that we were able to transport to different areas of the idea. With this design, we hoped to incorporate flow by recirculating water through several impeller pumps. We built a flow diffuser and baffles to disperse flow exiting the pump hose and reduce turbulent eddy formation. The final design was 4m x 1m with clear plastic walls to allow for light penetration (Fig. D.2.). We used marine batteries to power the pumps, however, we struggled to maintain power for more than 12 hours. Due to logistical constraints and long setup time, we decided to remove flow and treat these mesocosms as chambers, rather than flowing flumes. We ran experiments at five locations to test the effect of hydrogeomorphic zone and chronosequence on nitrate uptake. However, we ran into more problems with porewater leakage. This was particularly difficult at sites with many roots, which made inserting all four walls of the enclosure into the sediment very difficult. Despite small tides, the development of a pressure differential between inside and outside of the chambers created leakage through the sediment. We observed up to 20 cm of water level difference before equilibration began. We tested a tide equilibrator, which would expand and contract as water levels change to prevent development of water level difference. We tested this at multiple sites, but the increased complexity of mass balance calculations, as well as lack of flow from pumps, we decided to simplify our design. We switched our design to the circular mesocosms as detailed in Chapter 4. While logistics were still complex with providing multiple days of power, we saw reduced porewater leakage, especially as we were able to insert the mesocosms at least 20 cm into the sediment.
Figure C.2. Portable field flume with flow diffuser, baffles, and pumps for creating flow.

Figure C.3. Relative conductivity changes for each short-term experiment
Figure C.4. Ammonium fluxes during the short-term pulse experiments

Figure C.5. Chlorophyll concentrations over the course of the short-term pulse experiments
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Vita

Alexandra Christensen grew up in the hot desert of Arizona where she developed a love of the outdoors. While studying at the University of California, Berkeley, she became interested in wetland ecology and completed her Bachelor of Science in Conservation and Resources Studies. She worked and volunteered in several wetland and soil research labs and discovered that even Arizona has wetlands. After living in her car for 6 months while traveling around the country, she decided to make Baton Rouge her next home and entered the Coastal and Ecological Engineering Master’s program under Clint Willson and Robert Twilley. She simultaneously pursued her Ph.D. in Oceanography and Coastal Sciences, which culminated in the dissertation above. She is expected to graduate in May 2020 and will be starting a post-doctoral position at the Jet Propulsion Laboratory. When not doing field work or running samples in the lab, she can be spotted kayaking through the swamps or climbing in the mountains.